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THE UNIVERSITY OF ALBERTA

EVAPORATION FROM LAKE WABAMUN, ALBERTA

by

© ANDREJ SAULESLEJA

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

IN

METEOROLOGY

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA

FALL, 1976

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled "Evaporation from Lake Wabamun, Alberta," submitted by Andrej Saulesleja in partial fulfilment of the requirements for the degree of Master of Science in Meteorology.

Keith A. Hage
Supervisor

E. Reinelt

Meyer Lutzen

Date AUGUST 9, 1976

DEDICATION

To Eric and Peter, my sons

and to my loving wife, Glenda

ABSTRACT

Evaporation from Lake Wabamun, Alberta is estimated over a period of 78 days during the summer of 1974 using hourly data in a mass-transfer equation of the form:

$$E = \rho C_E (u - u_s)(q - q_s)$$

It is shown that a wind-and-stability-dependent formulation for C_E provides an estimate for evaporation which does not differ significantly from evaporation calculated using a constant value for $C_E = 0.0016$. The evaporation calculated using the mass-transfer method is found to be consistent with lake evaporation estimated by applying a simplified water budget to the lake.

The use of data in the form of daily averages of wind, temperatures, and specific humidities in the mass-transfer equation yielded a 10% underestimate of evaporation over the period of study.

Upper and lower bounds to enhanced evaporation from Lake Wabamun due to the presence of heated plumes from power plants are formulated, and it is found that slightly less than one half of the excess thermal pollutant goes into

evaporating water, while the remaining energy is distributed between heat loss due to radiation and sensible heat transfer. It is estimated that the power plants operating at a capacity of 900 MW would increase annual evaporation over the lake by approximately 2 cm. The enhanced evaporation is found to be less than the computed mean annual discharge (6 cm) from Wabamun Creek.

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CHAPTER I

INTRODUCTION

1.1 Background for the Study

Lake Wabamun, Alberta has been the site of environmental impact studies in recent years. The lake is within an hour's drive of Edmonton, and serves that centre and many towns and districts as a recreation area (Fig 1). In addition to summer cottages, beaches, the Paul Band reserve, and a provincial park bordering its shores, there are two power plants located at the eastern end of the lake. The power stations use the lake water for coolant, and inject heated plumes of water onto the lake surface. These plumes have been a concern of the environmental sciences for a number of years. Along with claims of accelerated weed growth and fish kills, voiced by local residents, there has also been some suspicion that, in time, the increased evaporation resulting from the heated water would lower water levels in the lake.

Hogg (1973) using fast-response instrumentation compared evaporation over the hot plume, and over the unaffected lake surface, found greatly enhanced evaporation over the heated water. Hogg could not, however, make firm

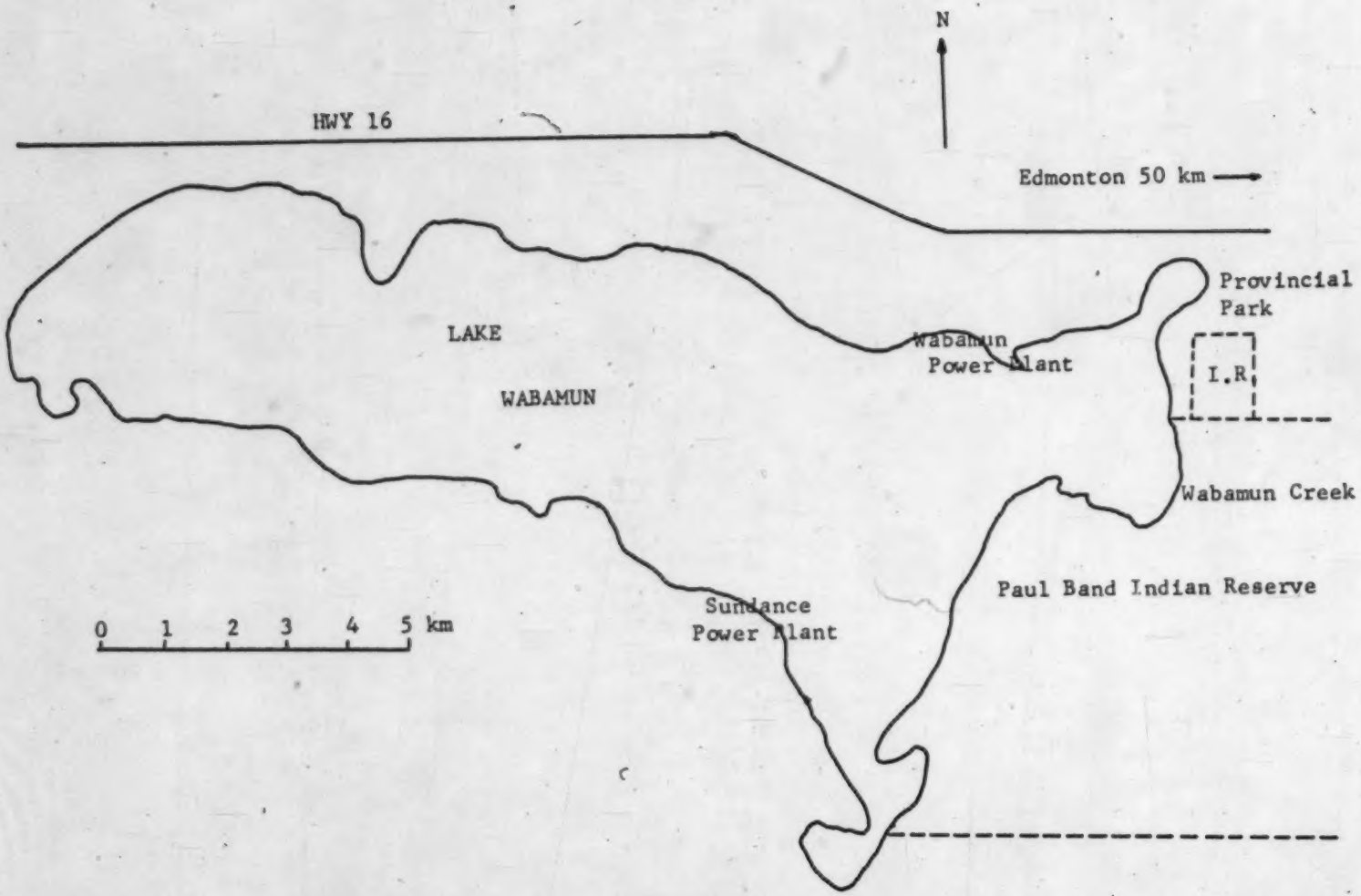


Figure 1. Lake Wabamun, Alberta.

conclusions about the long-term effects, since his findings were based on two warm days in August and a cold day in October.

In response to public concern, Calgary Power Limited has constructed a large cooling pond and has diverted North Saskatchewan River water to service the Sundance plant on the south shore of the lake. In addition, water levels at Lake Wabamun in 1974, as in many Alberta lakes that year, were extraordinarily high and public concern for enhanced evaporation has waned somewhat.

The studies of Fritz and Krouse (1973), and Nursall, Nuttall and Fritz (1972) of the isotopic constituents and salt content of Lake Wabamun indicate that the lake may have a significant throughflow and inflow of groundwater. Their hypothesis is supported by Carlson (1970) who suggests that a buried Pleistocene riverbed connects Lake Wabamun with Lake Isle, and drains some of this water to the south east. Their thesis is that this aquifer both charges, and discharges water from the lake. They further estimate that there is a net inflow of groundwater into the lake which would be sufficient to raise the water level of the lake by 3-6 cm monthly in the absence of evaporation and rainfall

It was the hope of this study that, with data collected during 1973 and 1974 from a meteorological station network on the shores of Lake Wabamun, the effects of the heated plumes on evaporation, and the net flow of ground water into

the lake could be calculated. No comparable studies involving meteorological lakeshore measurements have been made for any Alberta lake. All prior estimates for lake evaporation in Alberta were based on data from existing climatological and synoptic observing networks.

1.2 The Choice of Procedures

Many methods are available for the study of mass and heat transfers . A battery of fast-response instrumentation could provide a good and reasonably representative estimate of lake evaporation by eddy-correlation methods, but the resources of equipment, and the labour required for such a study are prohibitive. Atmometers such as evaporation pans serve as analogues to lake conditions, and thrcugh the use of formulas such as those of Kohler et al. (1955) and others produce reasonable estimates of lake evaporation from shallow lakes and reservoirs. The mass-transfer approach for estimation of evaporation (Jacobs 1942 and others) was used in this study, because it provides a direct check of Hage's (1974) calculation of Lake Wabamun evaporation based on monthly averages of meteorological variables. The mass-transfer approach has an advantageously low cost in terms of labour and instrumentation compared to other methods.

Intercomparisons of lake evaporation calculated by differing methods were made by McKay and Stichling (1961).

Comparisons were made of evaporation estimated by the formulas of Meyer(1942), Penman(1948), Marciano and Harbeck(1954), and Kuzmin(1951) with evaporation estimates based on evaporation-pan, floating-pan evaporation, and water-budget measurements for Weyburn Reservoir, Saskatchewan. Large discrepancies were observed between the various estimates of evaporation. It was pointed out, however, by Harbeck(1961) that much of the observed discrepancy could have arisen from the manner in which the parameters of wind and water temperature were obtained for this study. Buckler(1973) estimated evaporation rates for Lake Diefenbaker, Saskatchewan, a large deep lake, and concluded that stored heat played an important role in the evaporation process in the late fall. Spring and Schaefer (1974) used a modified form of the Lake Hefner mass-transfer equation to obtain daily estimates of evaporation for Babine Lake in Northern British Columbia. Recently, McBean (1975) computed fluxes of heat and moisture from Lake Ontario during a cold frontal passage using a mass-transfer equation to compute evaporation. A significant feature of McBean's calculations was the use of a bulk-transfer coefficient which varied with the wind speed.

The Monin-Obukhov Similarity Theory (Lumley and Panofsky 1964) implies that the non-dimensional bulk-transfer coefficients for momentum,heat and water vapour are equal to each other at the limit of neutral stability. McBean (1975) used the results of Smith and Banke's (1975)

estimates of $(C_D)_N$ as estimates for C_E to compute evaporation from Lake Ontario. Hicks (1972) evaluated the bulk-transfer coefficients for momentum, water vapour and sensible heat over water bodies of different sizes and found that, while drag coefficients increase slightly with wind speed at moderate to high winds, at low wind speeds they do not differ significantly from those over aerodynamically smooth surfaces. Hicks also found an increase of the drag coefficient with increasing instability.

After considering the constraints of available data, it was decided that a bulk-transfer approach would be used along with water-budget estimates of evaporation in an attempt to determine whether or not there exists a net groundwater inflow into the lake, and to estimate the effects of the thermal effluent from the two power stations.

The formulation of the bulk-transfer coefficient (Jacobs 1942) used by Hage (1974) to estimate evaporation from Lake Wabamun made no allowances for variations with wind speed or stability of this coefficient. In this thesis the theoretical results of Deardorff's (1968) calculation of drag coefficient variation with stability is used to adjust the transfer coefficient for varying stabilities over the lake. The wind-speed-dependent formulation for the neutral stability drag coefficient used in this thesis was based on Charnock's (1954) dimensional analysis argument, and the observations of Smith and Banke (1975), and Wu (1969) for

high wind speeds. Nikuradse's (1933) profile formulation specifies a bulk-transfer coefficient corresponding to aerodynamic smoothness for low wind speeds. This formulation for the drag coefficient is in agreement with some of the results of Hicks (1972). Chapter III of this thesis expands on the reasons for these choices.

1.3 The Experiment

There are few experiments which are carried out under ideal conditions, and this is not an exception. At the outset it was hoped to include data collected in the summer of 1973, as well as 1974 in the observations. However, reduction of the wind observations for 1973 was not completed in time for inclusion in this analysis. The conclusions found here were based on 78 days of data collected during the period 11 June 1974 to 28 August 1974. While the data were restricted to the summer months, some estimates were made of the effects of the thermal plumes during the winter months on lake evaporation, and an estimate was made of the magnitude of the yearly evaporation from the plume using the bulk-transfer method.

CHAPTER II

DATA COLLECTION

2.1 The Observing Sites

Lake Wabamun (Fig. 1), located 67 km west of Edmonton, is a large eutrophic lake surrounded by gently rolling parkland. The longest axis of this lake extends from WNW to ESE more or less along the directions of the prevailing winds. The morphometry of Lake Wabamun is outlined in Table 1.

The power stations and the observing sites are located at the eastern end of the lake (Fig 2). A thermohygrograph in a Stevenson screen, a Ryan water-temperature thermograph, a cup anemometer, a Stevens water-level recorder, and a staff gauge were installed on the north side of the lake, at the end of the Alberta Concrete Products Limited berm (Table 2 and Fig 3). A thermohygrograph in a Stevenson screen, two cup anemometers, a Stevens water-level recorder, and a staff gauge were installed at the end of the intake canal for the Sundance Power Plant, on the south side of the lake (Table 2 and Fig 4).

The topography of the shores adjacent to the instrument sites is shown in Fig 2. The steep slopes to the east of the Alberta Concrete Products Ltd. berm dictate that only the westerly to southeasterly wind directions off the lake

TABLE 1

Morphometry of Lake Wabamun
(after Hogg 1973)

Elevation	722.7 m
Area	82.5 km^2
Volume	ca. 0.455 km^3
Length	19.2 km
Maximum breadth	6.6 km
Mean breadth	4.3 km
Maximum depth	11.6 m
Shoreline length	57.3 km
Area of surface drainage	372.4 km^2

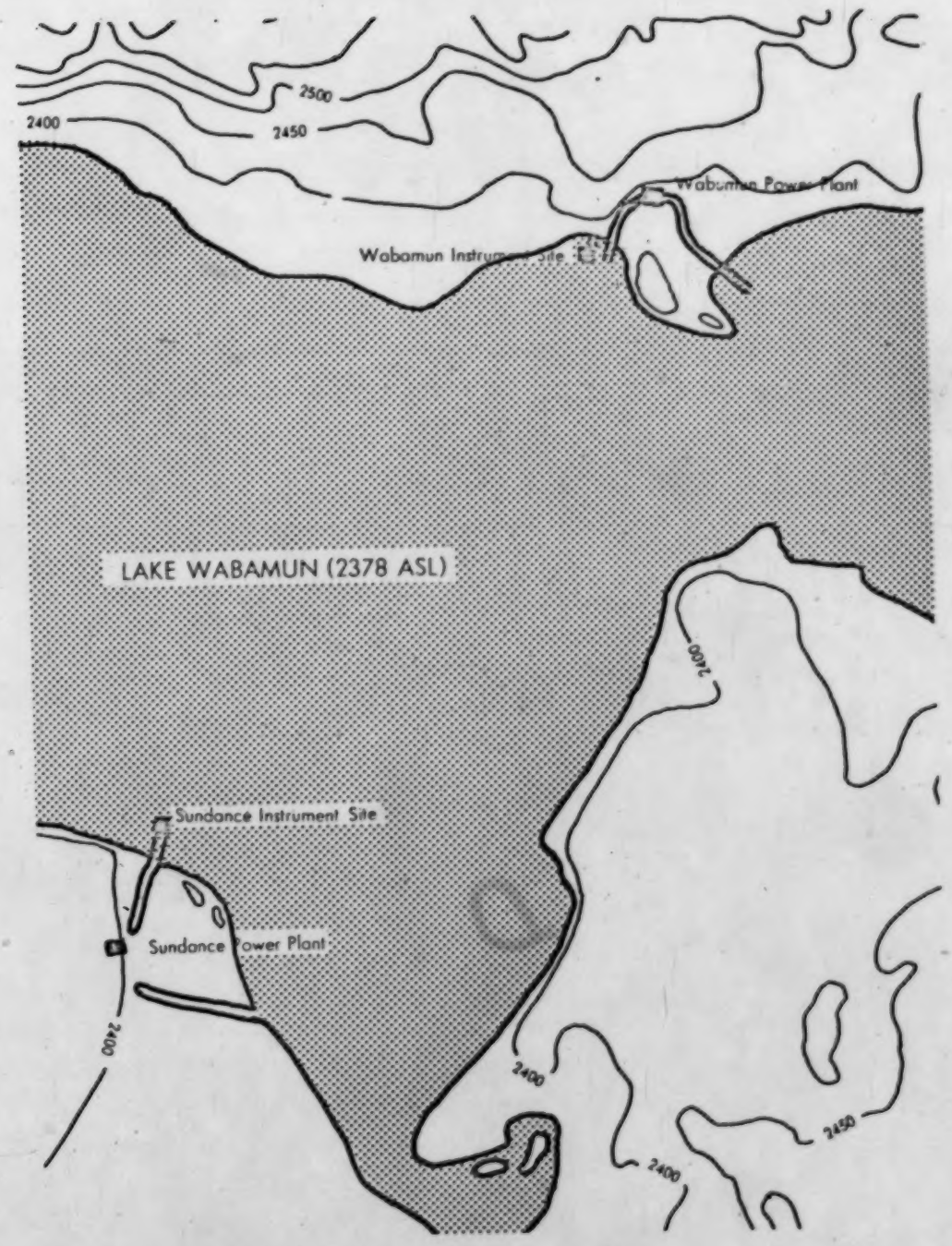


FIGURE 2. Lake Wabamun Observing sites and Topography.

TABLE 2

Instrumentation

Location	Element	Height (28/8/74)	Sensor
Sundance	Wind speed	3.67 m	Cup anemometer
	Wind direction	3.67 m	Vane
	Air temperature	1.54 m	Thermohygrograph
	Relative humidity	1.54 m	Thermohygrograph
	waterlevel		Stevens recorder
Wabamun	Wind speed	3.12 m	Cup anemometer
	Wind direction	3.12 m	Vane
	Air temperature	1.05 m	Thermohygrograph
	Relative humidity	1.05 m	Thermohygrograph
	waterlevel		Stevens recorder
	Water temperature	-6 cm	Ryan Thermograph

FIGURE 3. Pictures of the Wabamun Observing Site.

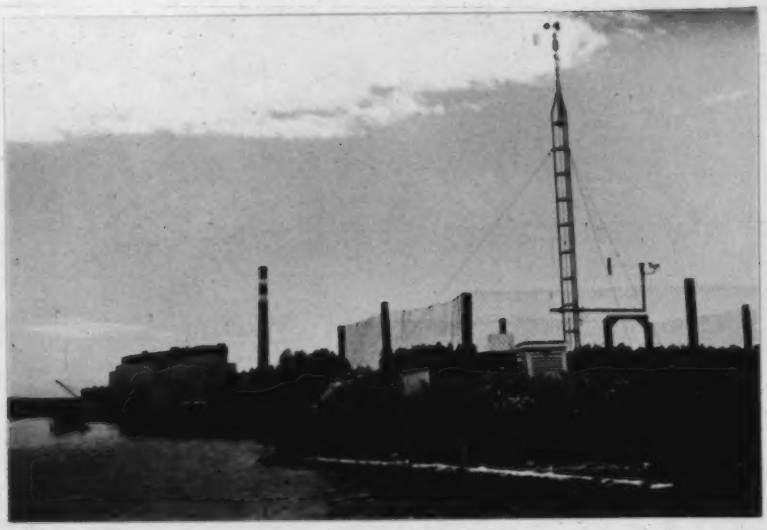


FIGURE 4. Pictures of the Sundance Observing Site.

could be used for mass-transfer calculations. The topography immediately around the Sundance site was fairly flat, however.

2.2 Instrumentation and Calibration

The instruments (Table 2, Figs 3 and 4) were inspected, and calibration measurements were taken at approximately weekly intervals during the summer of 1974 by Mr. Bob Weatherburn and Dr. John Honsaker of the Meteorology Divison at the University of Alberta. The cup anemometers were calibrated in the Department of Civil Engineering wind tunnel prior to, and after the experiments. The procedures followed for each instrument are outlined below. Because of the magnitude of error involved, the Assman psychrometer used for field calibrations also required recalibration.

2.2.1 Assman Psychrometer

As a calibrating instrument, the Weather Measure Corp. Assman psychrometer model H 331 was a disappointment. When a check was made in the laboratory of the readings of the wet and dry bulb thermometers used for calibrating the thermohygrographs, it was found that a discrepancy of 0.8 C existed between the wet and dry-bulb thermometers. The calibration of the Assman psychrometer was performed by placing both the wet-and dry-bulb thermometers initially in an ice bath and then, after the ice melted, observing the

temperatures as the water temperature rose to room temperature. By subtracting 0.8 C from all the dry-bulb thermometer readings used in calibration of the thermohygrographs, the standard deviation between the wet and dry-bulb thermometers was reduced to 0.2 C.

2.2.2 Thermohygrographs

The thermohygrographs were initially adjusted, and then cycled through changes in temperature and relative humidity within the environmental chambers of the Department of Botany of the University of Alberta. The calibrations applied to these instruments were obtained from weekly comparisons with the Assman psychrometer.

Least-squares regressions of corrections of the form:

$$T_{true} = C_1 T_{measured} + C_2 \qquad (2.1)$$

$$R_{H_{true}} = C_1' R_{H_{measured}} + C_2' \qquad (2.2)$$

were made to the observations recorded by the thermohygrographs. The corrected relative humidity and temperature were plotted against time, relative humidity, and temperature to check for any remaining trends. No obvious trends existed, and there appeared to be no deterioration or aging of the hygrograph elements during the period in which the experiment was conducted. The

thermohygrographs were mounted within an unaspirated Stevenson screen, and thus there is the possibility of a considerable wind-speed-dependent time lag between the conditions in the environment, and screen measurements. Bryant (1968) examined the time-lag for the Stevenson screen and found lag times to vary from 30 min in calm conditions to 6.5 min with 7 m/s winds. Calibration measurements taken with the Assman psychrometer would be the instantaneous conditions of the environment. No compensation was made for lag.

2.2.3 Ryan Water Temperature Thermograph

The Ryan water-temperature thermograph located at the Wabamun site was calibrated against periodic checks of water temperature using a standard thermometer. After addition of 1.7 C, the Ryan temperatures agreed to within +0.9 C with the calibrating thermometers.

There were no obvious instances where it could be definitely concluded that the hot-water plume from the Wabamun power plant was influencing the water temperature readings.

During the 78 days of the experiment water-temperature data were unavailable during the periods 0000 MDT, 29 June to 1700 MDT, 2 July and 0600 MDT, 5 August to 1400 MDT, 6 Aug.

2.2.4 Stevens Water-level Recorder and Staff Gauges

Except for short gaps caused by jammed charts etc. At the Sundance site, a nearly continuous record of lake levels was available both at Wabamun and Sundance instrument sites. The charts in the Stevens recorders were changed, and a reading was taken of the water level at each site from the staff-gauge at weekly intervals. Calibrations factors based on the site staff-gauge measurements were then applied to data extracted from each site. The calibration calculations resulted in a record of the height above water of the thermohygrograph units, used later in this work for mass-transfer calculations.

The water-level measurements of the two Stevens recorders provide a continuous record of water-level observations. These records were combined to produce lake levels above mean sea level using water-level measurements made by Nuttal (1974). It was also necessary to compensate for wind setup in the water-level readings.

2.2.5 Anemometers

A continuous record of wind speed and direction was available from the strip-chart record of the Rustrak event recorders. Readings from the strip-chart record were tabulated on magnetic tape, corrected for time and reduced to hourly means of calibrated wind speeds using a minicomputer. The anemometers and their recorders were

calibrated in the Department of Mechanical Engineering wind tunnel using the procedures described by Honsaker (1975).

2.3 The Local Weather Conditions During the Experiment

The unusually heavy snows of the winter of 1973-1974[1] helped to produce extraordinarily high water levels in nearby lakes and sloughs during the spring and summer of 1974. Water levels were high enough to produce a large surface discharge from Wabamun Creek, at the east end of Lake Wabamun.

The summer of 1974, or at least that portion within the 78 days of the study, could be described for the most part as showery, and cool. On 32 of the 78 days measurable precipitation was recorded by the rain gauges. Except for localized or short-duration winds during storms, the monthly mean winds were lower than normal. Relevant climatological statistics[2] for June, July, and August 1974 are given in the monthly summaries from the Atmospheric Environment Service below:

June

> The month of June in Edmonton was mainly sunny, warm and contrary to month end statistics, was generally dry.

---

[1] Snowfall recorded at the Edmonton International Airport was 219.5 cm vs 132 cm normally

[2] Since the summaries were written before metrication the units are non-metric.

Sunshine, with 305 recorded hours, was 40 hours above normal and the sunniest since 1969.

The mean daily temperature of 61.7 degrees was 3 degrees above normal. Temperatures ranged from a low of 44 recorded on several days to a high of 88 degrees on the 23rd.

Total precipitation of 4.87 inches for the month was much above normal. Of this amount, 3.15 inches or 0.21 inches above the normal total of 2.94 inches for the entire month fell on June 26. New 6, 12 and 24 hour rainfall amounts for June were established on this date.

Thunder recorded on 10 days, was above the normal of 4.4 thunderstorm days for the month.

July

Temperatures during the month of July in Edmonton were slightly below normal during the first half but recovered to near or slightly above normal during the later half.

Precipitation for the month totaled 4.97 inches, 1.69 inches above normal. Of this total, 4.26 inches fell during 10 of the first 12 days of the month.

Two significant storms passed through the area. A rain storm during the 11th-12th accompanied by winds gusting to 51 mph dumped 2.24 inches of rain over the city resulting in localized flooding for the 2nd time in 16 days. A vicious thunderstorm crossed the northern portion of the city during the late afternoon of the 30th. Winds at the Industrial Airport reached gusts of 47 mph however, near the storm's center, Namao reported gusts to 87 mph. Apart from these two windstorms average wind speed for the month was only 7.8 mph, the previous low average speed set in 1958.

August

The month of August in Edmonton was cool, dry and cloudier than usual. With a mean daily temperature of 58.5 degrees, 2 degrees below normal, the month was the coolest August since

1968. Temperatures reached into the 80's on only 2 days. Those were on August 3 and 4 when the daily maximum temperatures were recorded as 85 and 86 degrees respectively. The lowest temperature for the month was 42 degrees on the 13th.

Precipitation for the month was 1.67 inches below normal and the first month since July 1973 that total precipitation was not above normal.

Sunshine with 252.7 hours recorded was 16 hours below the long term normal. Winds during the month were lighter than average. Highest gust was from the northwest at 45 mph on the 24th.

CHAPTER III

THEORY

3.1 Introducing the Assumptions

3.1.1 The Constant-Flux Layer

Most profile and mass-transfer theories used in micrometeorology are derived by considering an infinite homogeneous level surface. For a lake the size of Wabamun, such an approach is justified. Because the lake is small compared to synoptic-scale systems and because the layer of interest is shallow, pressure-gradient and Coriolis forces may be neglected in the equations of motion. Since turbulent diffusion is much greater than molecular diffusion in the free atmosphere, molecular diffusion is neglected also. Furthermore, radiative flux divergence will be small except in cases of extremely stable stratification (Lumley and Panofsky, 1964). What remains is a constant-flux layer near the ground. With few exceptions, the stability regime over the lake favours lapse conditions The assumption of a constant-flux layer may not be fully justified over the hot-water plume, where horizontal gradients of water-surface temperature exist.

In summary, it is assumed that the earth's rotation, radiative flux divergence, and molecular diffusion, are

negligible. With the futher assumptions of steady state, horizontal homogeneity, and the Boussinesq approximation (density a function of geopotential) one obtains the following equation of motion for the mean flow:

$$\frac{d\vec{v}}{dt} = \frac{1}{\rho}\frac{\partial \vec{\tau}}{\partial z} \tag{3.1}$$

3.1.2 The Monin-Obukhov Similarity Theory and More Assumptions

The Kernel, or starting point for the transfer equations used in this thesis is the Monin-Obukhov similarity theory. By defining the Monin-Obukhov length:

$$L = \frac{-u_*^3}{k\left(\frac{g}{T_v}\right)\left(\langle w'T_v'\rangle\right)} \tag{3.2}$$

where:

$$\langle w'T_v'\rangle = \langle w'T'\rangle + 0.61\,T\langle w'q'\rangle$$

One may then represent the profiles of wind, temperature and humidity as functions of z/L:

$$\frac{kz}{u_*}\frac{\partial u}{\partial z} = \phi_1\left(\frac{z}{L}\right) \tag{3.3}$$

$$\frac{k u_* z}{\left(\frac{-H}{\rho C_p}\right)}\frac{\partial \theta}{\partial z} = \phi_2\left(\frac{z}{L}\right) \tag{3.4}$$

$$\frac{k u_* z}{(-E/\rho)} \frac{\partial q}{\partial z} = \phi_3 \left(\frac{z}{L}\right) \tag{3.5}$$

The bulk-transfer coefficients for momentum, heat and water vapour are defined as:

$$u_*^2 \equiv \frac{\tau}{\rho} = C_D (u - u_s)^2 \tag{3.6}$$

$$-\frac{H}{\rho C_p} = C_H (u - u_s)(\theta - \theta_s) \tag{3.7}$$

$$-\frac{E}{\rho} = C_E (u - u_s)(q - q_s) \tag{3.8}$$

In order to remain consistent with the equations derived by Deardorff (1968) for the bulk-transfer coefficients, a further assumption needs to be made. Deardorff has assumed that $(C_D)_N = (C_H)_N = (C_E)_N$. This is equivalent to the assumption that $\phi_1 = \phi_2 = \phi_3$ $\lim \frac{z}{L} \to 0$, and that the roughness lengths for momentum, moisture and heat are equal. These assumptions are equivalent to considering the transport of momentum by pressure fluctuations to be negligible. Calder (1966) has reviewed the Monin-Obukhov similarity theory, and the role of the pressure fluctuations. Businger et al. (1971) have obtained $K_H/K_M = 1.35$ and $k = 0.35$. These results would seem to indicate that this transport term should not be neglected. However, Dyer (1974) argues that the 10% reduction in wind

speed and 33% reduction in drag-plate shear-stress which Businger et al. (1971) used to calibrate their instruments was too large, and that lower values would result in $K_w/K_m = 1.0$ with k=0.4. The controversy concerning the role of pressure fluctuations, and momentum transport continues.

Another complication which is introduced if one considers momentum transport due to pressure fluctuations is that there should be different roughness length scales for momentum, heat and water vapour. Brutsaert (1975) has developed equations for the roughness lengths for heat and water vapour if the momentum roughness length is known. The theory developed is useful for $Re_* < 0.13$ and $Re_* > 2$, where Re is the interfacial Reynolds number $Re = \frac{u_* z_o}{\nu}$. Interfacial Reynolds numbers found over a water surface include the range $0.13 < Re_* < 2$ where no useful theory exists.

Since there is no solid theory to use as a guide when estimating the roughness lengths for mass, moisture and heat, it was assumed that these were equal. Furthermore, pressure transport of momentum is assumed to be negligible. These assumptions are consistent with the bulk-transfer coefficient formulations of Deardorff. For consistency, k=0.4.

3.2 The Formulation of the Bulk-Transfer Coefficients.

3.2.1 Introduction

Experimental estimates of the bulk-transfer coefficient for evaporation have been obtained by various methods. Hage (1974) has outlined a number of estimates (Table 3), and a wide variation is apparent among the values for the transfer coefficient. Estimates of C_e range from 1 to 2.3. Hage argues that the value of 0.0016 for C_e, the Lake Hefner result, is probably the most realistic for Lake Wabamun, since it would best represent the unstable conditions present over the lake surface.

Another route to obtaining a value for the bulk-transfer coefficient is to integrate the profiles of temperature, moisture, and momentum based on the similarity theory of Monin and Obukhov, and others. Via this route one may introduce a dependence on wind, and thermal stratification, into the bulk-transfer coefficient formulation. By defining a bulk-transfer coefficient in this manner one should be able to apply the result to any wind and stability condition, and obtain meaningful results over the heated plumes also.

The following pages contain the steps by which a formulation of the bulk-transfer coefficient based on profile theory is obtained (Deardorff 1968).

TABLE 3

Some Estimates for the Bulk-Transfer Coefficient for Evaporation (after Hage 1974)

Source	Remarks	C_E (x 10^3)
Jacobs (1942,1943)	Oceans, indirect	2.3
Privet (1960)	Oceans, indirect	1.8
Budyko, et al.	Oceans indirect	1.8
Wust (Sverdrup, 1951)	Ocean, indirect, pan	1.1-1.4
Bunker (1960)	Ocean, indirect, aircraft	1
Deacon and Webb (1962)	Lakes	1.1-1.6
Pond, et al. (1971)	BOMEX, eddy correlation BOMEX, dissipation	1.23+.17 1.25+.25
Marciano and Harbeck (1954)	Lake Hefner, Water Budget	1.6
Linacre et al. (1970)	Lake Wyanagan, eddy correlation	1.4
Hicks,1971	Lake Michigan, eddy correlation	1.5

3.2.2 The Equivalence of the Bulk-Transfer Coefficients at Neutral Stability

In the limit of neutral stability, as $\frac{z}{L} \to 0$, $C_D = C_H = C_E$, a result of previous assumptions.

Accordingly, by integrating (3.3) one obtains:

$$\frac{u - u_s}{u_*} = \frac{1}{k} \ln \frac{z}{z_0} = \frac{1}{\sqrt{C_D}} \qquad (3.9)$$

Integrating (3.5) one obtains:

$$(\theta - \theta_s) \frac{u_*}{(-H/\rho C_p)} = \frac{1}{k} \ln \frac{z}{z_q} \qquad (3.10)$$

Again assuming a common roughness for momentun, heat and moisture.

$$(\theta - \theta_s) \frac{u_*}{(-H/\rho C_p)} = \frac{1}{\sqrt{C_D}}$$

$$(z_q = z_0)$$

or, $$(\theta - \theta_s)(u - u_s) C_D = (-H/\rho C_p)$$

$$\therefore \quad (C_D)_N = (C_H)_N$$

Similarly, for water vapour in the neutral limit:

$$(C_E)_N = (C_D)_N$$

Thus, for this formulation assumed by Deardorff (1968), $(C_E)_N = (C_H)_N = (C_D)_N$ and, for the neutral case, it is sufficient to know $(C_D)_N$ in order to know $(C_H)_N$ and $(C_E)_N$

also.

3.2.3 Extending the Theory to Non-Neutral Stability.

The stability of the atmosphere is described by the ratio z/L. If one combines (3.6) to (3.8) and (3.2) one gets:

$$\frac{z}{L} = \frac{k\, C_H / (C_D)_N}{(C_D)_N^{1/2}\left[C_D/(C_D)_N\right]^{3/2}} \left\{ \frac{g}{T_v} \frac{z}{(u-u_s)^2} \left[\theta - \theta_s + 0.61 T \frac{C_E}{C_H} (q - q_s) \right] \right\} \quad (3.11)$$

It has been calculated by Deardorff that

$$\frac{C_H / (C_D)_N}{(C_D)^{1/2}\left[C_D/(C_D)_N\right]^{3/2}}$$

is approximately unity for $-.02 < z/L < 0.02$.

If $$(R_i)_B \equiv \frac{g}{T_v} \frac{z}{(u-u_s)^2} \left[\theta - \theta_s + 0.61 T \frac{C_E}{C_H} (q - q_s) \right] \quad (3.12)$$

then

$$\frac{z}{L} \approx k\, (C_D)_N^{-1/2}\, (R_i)_B \quad (3.13)$$

3.2.3.1 Unstable Case

Assuming that $K_\theta = K_H$ in the unstable case, ie., that both potential temperature, and specific humidity are conservative in the absence of molecular diffusion, and using the profile formulation of Businger (1966):

$$\frac{k u_* z}{-E/\rho} \frac{\partial q}{\partial z} = \left(1 - \gamma \frac{z}{L}\right)^{-n} \quad (n = \tfrac{1}{2} \text{ if } K_\theta = K_H) \tag{3.14}$$

$$\frac{k u_* z}{(-H/\rho C_p)} \frac{\partial \theta}{\partial z} = \left(1 - \gamma \frac{z}{L}\right)^{-\frac{1}{2}} \tag{3.15}$$

$$\frac{k z}{u_*} \frac{\partial u}{\partial z} = \left(1 - \gamma \frac{z}{L}\right)^{-\frac{1}{2}} \tag{3.16}$$

These equations have been integrated by Paulson (1967) with the results:

$$C_D/(C_D)_N = 1 - \frac{(C_D)_N^{\frac{1}{2}}}{k}\left[\ln\left(\frac{1+x^2}{2}\right) + 2\ln\left(\frac{1+x}{2}\right) - 2\tan^{-1}x + \frac{\pi}{2}\right] \tag{3.17}$$

$$C_H/(C_H)_N = \left[\frac{C_D}{(C_D)_N}\right]^{\frac{1}{2}} \left[1 - \frac{2}{k}(C_D)_N^{\frac{1}{2}} \ln\left(\frac{1+x^2}{2}\right)\right]^{-1} \tag{3.18}$$

where

$$x = \left(1 - \gamma \frac{z}{L}\right)^{\frac{1}{4}}$$

Saunders (1964) made the same assumptions. Consequently $\frac{C_E}{(C_E)_N} = \frac{C_H}{(C_H)_N}$, ie., $K_E = K_H$ and $n = \frac{1}{2}$. Paulson (1967) suggests that $K_E = K_M$ and $n = \frac{1}{4}$ and then $C_E/(C_E)_N$ coincides with $C_D/(C_D)_N$. More recently, Dyer (1974) in a

review of profile relationships has suggested that $\frac{C_E}{(C_E)_N} = \frac{C_H}{(C_H)_N}$ and $n = \frac{1}{2}$.

3.2.3.2 Stable Case

In the case of a stable stratification, Deardorff has used the log-linear profile (Lumley and Panofsky, 1964, p117) to represent the momentum, temperature and humidity profiles.

$$\frac{kz}{u_*}\frac{\partial u}{\partial z} = 1 + \beta_u \frac{z}{L} \qquad (3.19)$$

$$\frac{k u_* z}{(-H/\rho c_p)}\frac{\partial \theta}{\partial z} = 1 + \beta_T \frac{z}{L} \qquad (3.20)$$

$$\frac{k u_* z}{(-E/\rho)}\frac{\partial q}{\partial z} = 1 + \beta_q \frac{z}{L} \qquad (3.21)$$

$\beta_u, \beta_T, \beta_q$ = constants.

Integration of (3.19) to (3.21) yields:

$$\frac{C_D}{(C_D)_N} = \left\{1 + \beta_u \frac{C_H}{(C_H)_N}\left[\frac{C_D}{(C_D)_N}\right]^{-\frac{3}{2}}(R_i)_B\right\}^{-2} \qquad (3.22)$$

$$\frac{C_H}{(C_H)_N} = \left[\frac{C_D}{(C_D)_N}\right]^{\frac{1}{2}}\left\{1 + \beta_T \frac{C_H}{(C_H)_N}\left[\frac{C_D}{(C_D)_N}\right]^{-\frac{3}{2}}(R_i)_B\right\}^{-2} \qquad (3.23)$$

(3.24)

It has been assumed that the neutral transfer coefficients are equal. Equations (3.22) and (3.23) yield a quadratic in $(C_D/(C_D)_N)^{1/2}$.

$$\left(1-\frac{\beta_u}{\beta_T}\right)\frac{C_D}{(C_D)_N}-2\left(1-\frac{\beta_u}{2\beta_T}\right)\left[\frac{C_D}{(C_D)_N}\right]^{1/2}+\left[1-\frac{\beta_u^2}{\beta_T}(R_i)_s\right]=0$$

Deardorff has solved this quadratic, and his curves for $\frac{C_D}{(C_D)_N}$, $\frac{C_H}{(C_H)_N}$ and $\frac{C_E}{(C_E)_N}$ appear in Fig. 5 for $\beta_u=7$, $\beta_T=11$ and $\beta_w=20$. Deardorff explains that the discontinuities in the slope of the curves at $(R_i)_s=0$ arise from the fact that $\beta_u=\gamma/4$ and $\beta_T=\gamma/2$, and comments that more accurate measurements and theory are needed at this value of Ri.

Dyer (1974) suggests that $\phi_1=\phi_2=\phi_3=1+5(z/L)$ and thus the β's would all be equal to 5. It is noted that the choice $\beta=5$ would give a better match for the slopes of the $\frac{C_D}{(C_D)_N}$ curve under neutral conditions. Furthermore, the results of Businger et al. (1971) would indicate that K_H/K_m is nearly unity under strong stability. Thus β_u and β_T should be approximately equal.

The formulas above are cumbersome to use, and because the constants in them are not known to any great accuracy, an approximate formula given by Deardorff was used for computations. In the stable region these formulas are:

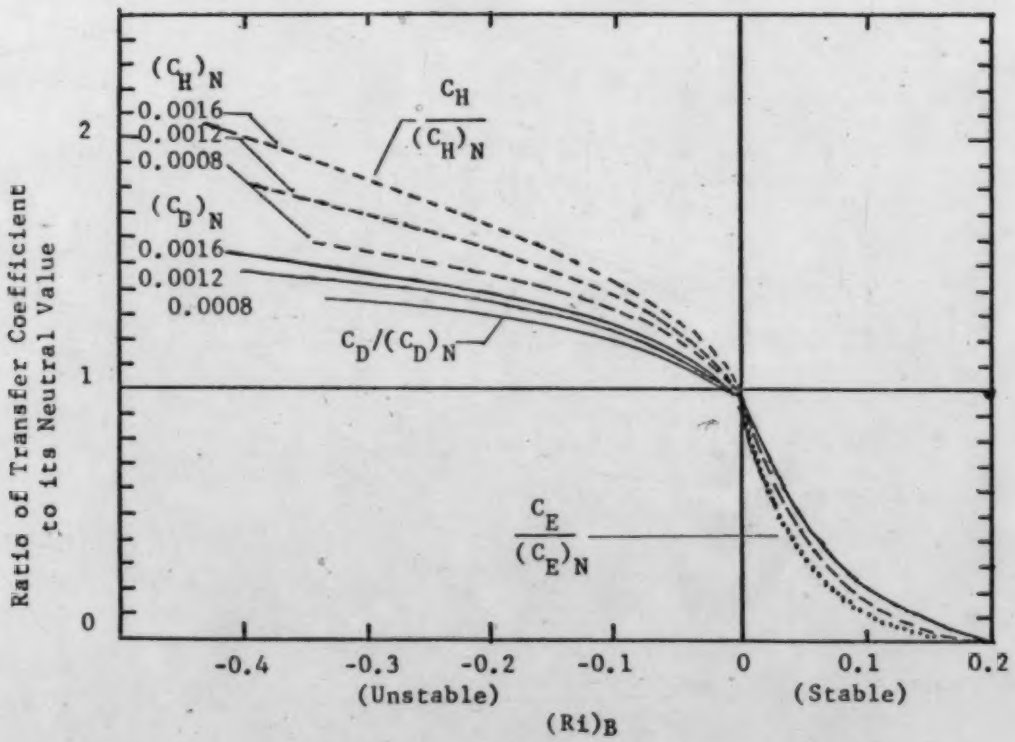


FIGURE 5. Ratios of the transfer coefficient to its neutral value from diabatic profile theory, over a wide range of Bulk-Richardson numbers (after Deardorff 1968).

$$\frac{C_D}{(C_D)_N} = \exp\left[-2\beta_u (R_i)_B\right] \tag{3.25}$$

$$\frac{C_H}{(C_H)_N} = \exp\left[-(\beta_u + \beta_T)(R_i)_B\right] \tag{3.26}$$

$$\frac{C_E}{(C_E)_N} = \exp\left[-(\beta_u + \beta_T)(R_i)_B\right] \tag{3.27}$$

with $\beta_u = 7$ $\beta_T = 11$ $\beta_T = 20$

In the unstable region,

$$\frac{C_D}{(C_D)_N} = 1 + \frac{7}{a} \ln\left[1 - a(R_i)_B\right] \tag{3.28}$$

$$\frac{C_H}{(C_H)_N} = 1 + \frac{11}{b} \ln\left[1 - b(R_i)_B\right] \tag{3.29}$$

where

$$a = 0.83 (C_D)_N^{-0.62}$$

$$b = 0.25 (C_D)_N^{-0.8}$$

assuming

$$\frac{C_E}{(C_E)_N} = \frac{C_H}{(C_H)_N}$$

With these formulas for one is now in a position to examine evaporation from a reservoir or lake with the effect of stability incorporated provided one has an estimate for

the bulk-transfer coefficient at neutral stability.

3.2.4 The Drag-Coefficient at the Limit of Neutral Stability

In order to apply the previously-developed theory, it is clear that an estimate for $(C_E)_N$, $(C_H)_N$ or $(C_D)_N$ is required. Near the neutral limit turbulence-induced variations in temperature become small and thus difficult to measure accurately. The wind fluctuations, however, remain measurable irrespective of the stratification, and thus $(C_D)_N$ should be easier to determine than $(C_H)_N$.

There are a number of recent estimates of the drag-coefficient at the near-neutral limit of stability (Table 3 and Fig 6). For wind speeds in excess of approximately 4 m/s, the formulas for C_D of Charnock (1955), Smith and Banke (1975), Hicks (1972), Shepherd et al. (1972) and Wu (1969) are in fairly close agreement. The equations are of the following forms:

(Charnock's equation)

$$\frac{u}{u_*} = \frac{1}{k} \ln \frac{gz}{u_*^2} + C \qquad \left(= \frac{1}{\sqrt{C_D}} \right) \tag{3.30}$$

where the constant C has been estimated as:

C=11.0 Charnock (1955)

C=10.4 Wu (1969)

C=10.3 Hicks (1972)

C=10.6 Smith and Banke (1975)

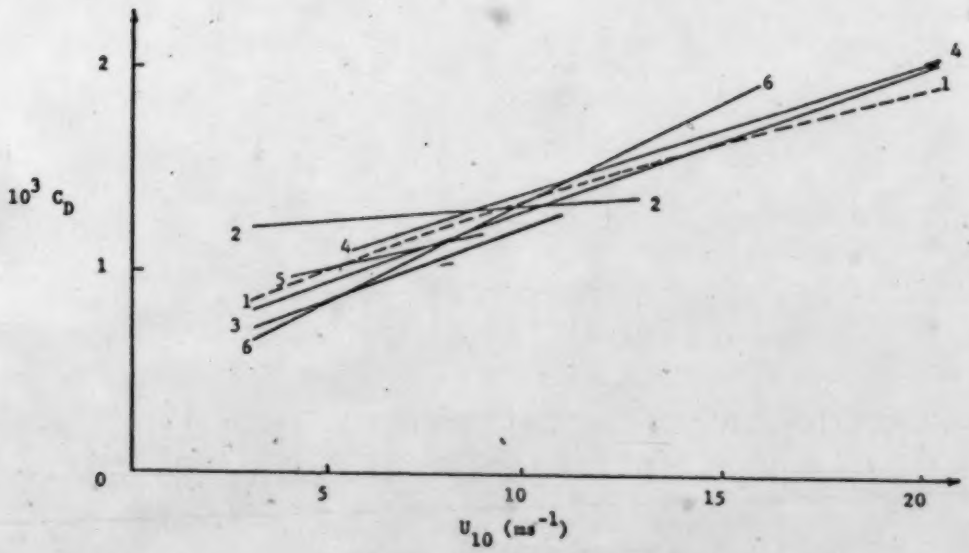


FIGURE 6. Drag coefficient and wind speed.

(1) Smith and Banke (1975):
(3.30) dashed curve
(3.31) line

(2) Brocks and Krugermeyer (1970)

(3) Hasse (1968)

(4) Miller (1964)

(5) Miyake et al (1972)

(6) Shepherd et al (1972)

(after Smith and Banke 1975)

$$10^3 C_D = 0.63 + 0.066 u_{10} \qquad (3.31)$$

Smith and Banke (1975)

$$10^3 C_D = 0.36 + 0.10 u_{10} \qquad (3.30)$$

Shepherd et al. (1972)

Charnock's equation was preferred to the others because it was based on a physical argument, and was readily transformable from one height to another. This equation has only one constant requiring determination. The Charnock equation was chosen to represent the drag-coefficient at wind speeds over 4 m/s, with a value of 11 for the undetermined constant. This value appears to produce a mean functional value for C_D between some of the observed values of Tables 3 and 4 and Fig 6.

At very low wind speeds, however, one would think that the roughness length should approach that for aerodynamically-smooth turbulent flow. The measurements of Hicks (1972) have indicated that this may be the case. It remains uncertain as to how the drag-coefficient should behave in the transition region from smooth-to-rough turbulent flow. Indeed, the flow over the waves may be transitional at all windspeeds, and the arguments of

Charnock may not be applicable there (Garrat and Hicks, 1973, Kitaigorodskii, 1968 and others).

In this thesis the drag-coefficient is assumed to remain continuous in transition from smooth-to-rough turbulent flow. At low wind speeds, the flow is represented by the profile equation:

$$\frac{u}{u_*} = \frac{1}{k} \ln \frac{u_* z}{\nu} + 5.5 \qquad (3.33)$$

And at higher wind speeds by:

$$\frac{u}{u_*} = \frac{1}{k} \ln \frac{g z}{u_*^2} + 11 \qquad (3.34)$$

The drag-coefficient is the greater of that given by (3.33) or (3.34) depending upon the wind speed. The variation of the drag-coefficient with wind speed appears in Fig. 7.

3.2.5 The Properties of the Water Surface

In (3.33) and (3.34) for the drag-coefficient, it is usually assumed that the surface drift velocity u_s is negligible, or at least very small when compared to the wind speed u. Hicks (1972) measured the surface drift, and found that $u_s \cong u_*$. This result was incorporated into the calculations to follow.

Wu (1975) has also examined the surface drift and has obtained values for u_s equal to 3 to 4% of u_{10} for fetches such as those of Lake Wabamun. There is little difference between the method of Wu and Hicks since $u_s = 0.03u_{10}$ to

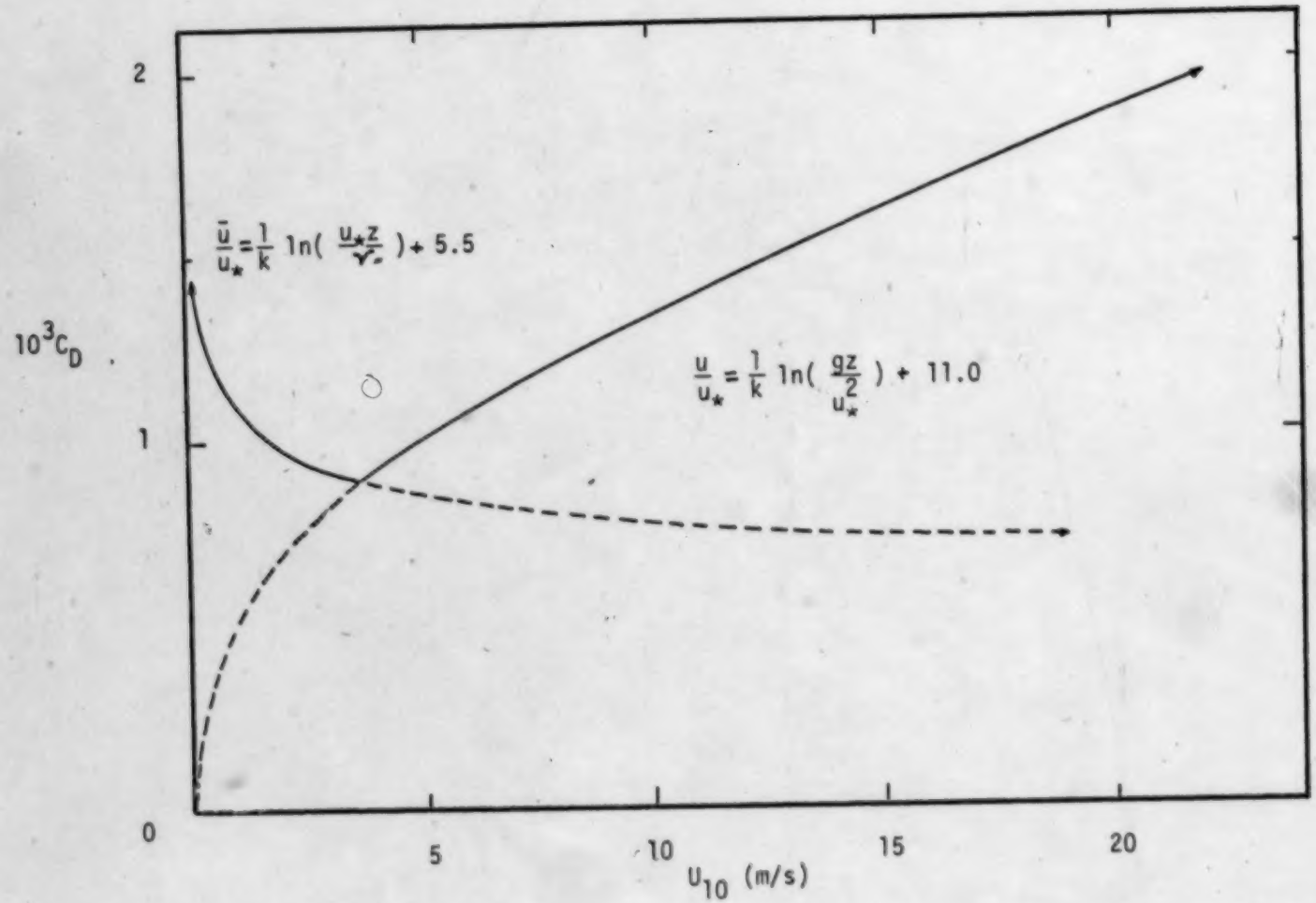


Figure 7. Selected drag-coefficients vs wind speed.

0.04u, and the error in choosing one over the other is but .5% of u which would result in a 1% error in estimating C_D.

3.2.6 The Method of Computation

The equations developed thus far are applicable to winds, temperatures and specific humidities measured at one level, and water temperature measured at the surface. Usually, however, winds and temperatures are not available at the same level, as is the case in this experiment. The winds were measured at 4 m above the water surface, and the temperature and relative humidity at 1 m to 1.5 m above the water. The water temperature was measured at a few centimeters below the water surface. This water temperature may be somewhat different than the "skin" temperature of the water. In the absence of solar radiation, a quantitative estimate has been made by Saunders (1967) of the temperature difference between "bucket" and actual surface temperatures. In this work, no correction was made for the difference between the "skin" or surface temperature and the temperatures measured some cm below the surface. The effect of solar radiation on the "skin" temperature is not clearly understood, and long-term measurements of radiation are not available. Furthermore, the Ryan thermograph was located near the shore, and, qualitatively, one would expect nocturnal water temperatures to be slightly lower than mid-lake temperatures, and daytime water temperatures to be slightly higher than mid-lake water temperatures.

Qualitatively then, the effects of site and depth of water temperature measurement would seem to cancel, and were thus ignored for calculation. The magnitude of the "skin-bucket" temperature difference is estimated by Saunders (1967) to be 0.3 C in the absence of radiation, and thus is a relatively small quantity.

In this study, the anemometer and thermohygrograph were located at different heights. Provided one uses the previously-mentioned profiles, one may obtain the following system of equations for parameters measured at levels 1 and 2:

$$(R_i)_B = \frac{g}{T_v} \frac{z_1}{(u_1-u_2)^2}\left[\theta_1 - \theta_2 + 0.61 T_1 \frac{c_s}{c_p}(q-q_s)\right] \qquad (3.35)$$

$$\frac{z_1}{L} = k (C_D)^{-1/2} (R_i)_B \qquad (3.36)$$

$$(C_D)_N = \begin{cases} \left(\frac{1}{k}\ln\frac{u_{z_2}}{\nu} + 5.5\right)^{-2} \\ \left(\frac{1}{k}\ln\frac{z_2}{u_{z_2}} + 11\right)^{-2} \end{cases} \qquad (3.37)$$

$$(u_2-u_1) = \int_{z_1}^{z_2} \frac{u_*}{kz} \begin{cases} 1+\beta_u \frac{z}{L} & \text{for } L>0 \\ (1-\gamma \frac{z}{L})^{-1/4} & \text{for } L<0 \end{cases} dz \qquad (3.38)$$

$$u_* = u_s \qquad (3.39)$$

Thus, assuming that q_s, T_1, q, u_2 are obtained from measurements, one is left with five equations in five unknowns u_1, u_*, u_s, L and $(C_D)_N$. Because an analytic solution to these equations did not appear possible, the system was solved numerically by successive approximations The subroutine for the solution of these equations appears in Appendix A. Hourly values of $C_D, C_E, L, u_1, (C_D)_N$ and the evaporation rate were obtained.

3.3 Other Methods

The method for evaporation calculation described previously is rather complicated and laborious. It would be preferable to use daily, weekly or monthly averages of temperature, wind and humidity to compute evaporation. For this reason the evaporation was recomputed using averages. Hage (1975) has examined the problem of computing evaporation using monthly mean values of temperature, relative humidity and wind in the bulk-transfer equation for evaporation. Hage expanded the saturation vapour pressure in a Taylor series using the Clausius-Clapeyron equation, and has modeled the diurnal variations in temperature,

relative humidity and wind speed to estimate covariance and non-linearity errors. During the summer months it was found that the model predicted that these errors would be less than about 5%. Hage found that averages could not be used during the winter months because wind and temperature were correlated because of the passage of synoptic-scale systems. An expansion like Hage's, of the stability-dependent formulation for bulk-transfer equation was not possible because the bulk-transfer coefficient is not monotonic with respect to wind speed. An estimate was made, however, of evaporation using averaged values of wind, temperature and humidity and a constant bulk-transfer coefficient.

3.4 Comparison of Bulk-Transfer Coefficients

The Lake Hefner experiments resulted in a mean value for the bulk-transfer coefficient of 0.0016. It was also possible to obtain a constant or weighted-average value for the stability-dependent bulk-transfer coefficient over a period of time.

Weighted averages are defined in the following manner:

$$\bar{C}_E \equiv \frac{\int_{t_1}^{t_2} C_E (u-u_s)(q-q_s)\,dt}{\int_{t_1}^{t_2} (u-u_s)(q-q_s)\,dt} = \frac{E\Big|_{t_1}^{t_2}}{\int_{t_1}^{t_2} (u-u_s)(q-q_s)\,dt} \qquad (3.40)$$

Such an average bulk-transfer coefficient may be defined for daily, monthly or larger time periods. For this study such

a mean was computed over a period of one day, and over all the days of measurement.

3.5 A Water-Budget Evaporation Estimate

The water-budget method was used to check the consistency of the results of evaporation calculations. The water budget for Lake Wabamun may be written:

$$L_{\ell} = I_{p} + I_{R} + I_{G} - O_{E} - O_{R} - O_{G} \qquad (3.41)$$

The runoff was estimated using Thornthwaite procedures described by Laycock (1967). The Thornthwaite procedure can not be expected to give good results if precipitation is distributed very unevenly over the month. Runoff was equated to surplus water by the following estimate of storage distribution for the Lake Wabamun basin:

$$I_{R} = 0.5S_{1/2} + 0.1S_{1} + 0.4S_{4} + 0.25S_{6} + 0.15S_{10} \qquad (3.41)$$

Where S_{x} represents surplus water at a storage capacity of x inches. The equation above is due to Professor A.H. Laycock, Department of Geography, University of Alberta, as are the monthly surplus estimates on which this calculation was based.

The net inflow of ground water as well as the net evaporation are obtained as residuals in the water-budget calculation. The groundwater inflow and outflow are assumed to be negligible for the purposes of calculating

evaporation. A large groundwater inflow would appear as an ever increasing discrepancy between evaporation calculated by the water-budget method and evaporation calculated by the mass-transfer equations. Laycock (1975) suggested that groundwater flow into Lake Wabamun is probably small compared with other terms in the water-budget.

3.6 Estimating Enhanced Evaporation

3.6.1 Introduction

Limits on the amount of enhanced evaporation due to the thermal discharge were obtained by considering two extreme cases of thermal effluent behavior. In the first instance it was assumed that the heated water was instantaneously mixed with the waters of the lake, producing a rise in overall lake temperature. In the second instance it was assumed that no mixing of the plume and lake water occured, and that the total excess energy of the plume was dissipated by radiation and turbulent transport of sensible and latent heat. Enhanced evaporation based on variable bulk-transfer coefficients and the previous considerations was calculated by the procedures outlined below.

3.6.2 Two Thermal-Effluent Models

To model the behaviour of the thermal effluent, the lake was assumed initially to be in equilibrium with its surroundings under steady-state conditions. It was also assumed that the heat flux, whether through conduction or infiltration of water through the bottom of the lake, is negligible. The energy budget may then be written as:

$$Q_R + Q_{LE} + Q_H = 0 \tag{3.43}$$

where:

$$Q_R = Q_{RS}\downarrow + Q_{RS}\uparrow + Q_{RL}\downarrow + Q_{RL}\uparrow \tag{3.44}$$

The Q_{RS} and Q_{RL} denote solar and terrestrial radiation, respectively, and the radiation is incoming or outgoing as indicated by the arrows. The outgoing terrestrial radiation $Q_{RL}\uparrow = \epsilon\sigma T_s^4$, but since $\epsilon = .97$ one approximates $\epsilon = 1$ in further calculations. With this approximation (3.44) may be written as:

$$Q_R = Q_{RS}\uparrow + Q_{RS}\downarrow + Q_{RL}\downarrow + \sigma T_s^4 \tag{3.45}$$

and

$$Q_H = -\rho C_p C_H (T_s - T_a)(u - u_s) \tag{3.46}$$

$$Q_{LE} = -\rho l_v C_E (q - q_s)(u - u_s) \tag{3.47}$$

where $$C_E = C_H \equiv C_{HE}\left(\frac{z}{L}\right) \tag{3.48}$$

If the rate at which heat is added to the water is Q per unit surface area, then the energy budget may be rewritten as:

$$Q_R' + Q_{LE}' + Q_H' = Q \tag{3.49}$$

With such an added input of heat the water surface temperature would rise by an amount ΔT_s. Assuming for the sake of this approximation that the temperature and specific humidity of the air, and the wind speed remain unchanged from unperturbed conditions over this surface, then one may write:

$$Q_R' = Q_{RS}\downarrow + Q_{RS}\uparrow + Q_{RL}\downarrow + Q_{RL}'\uparrow \tag{3.50}$$

where $$Q_{RL}' = \sigma (T_s + \Delta T_s)^4 \tag{3.51}$$

$$Q_{LE} = -\rho l_v C_E' [q_s' - q_a](u - u_s) \tag{3.52}$$

$$Q_H = -\rho C_p C_H' (T_s + \Delta T_s - T_a)(u - u_s) \tag{3.53}$$

and $$C_E' = C_H' \cong C_{HE}\left(\frac{z}{L}\right) \tag{3.54}$$

Combining the equations above one obtains:

$$\begin{aligned} Q = {} & \sigma[(T_s + \Delta T_s)^4 - T_s^4] \\ & + \rho l_v (u - u_s)\left\{ C_{HE}'[q_s' - q_a] - C_{HE}[q_s - q_a] \right\} \\ & + \rho C_p (u - u_s)\left\{ C_{HE}'[T_s + \Delta T_s - T_a] - C_{HE}[T_s - T_a] \right\} \end{aligned} \tag{3.55}$$

For the case of the thermal effluent diffusing throughout the whole lake, a further simplification is possible. Since ΔT_s is very small compared to T_s, and $C_{HE} \cong C_{HE}'$ one can expand the specific humidity and radiation using first-order approximations with respect to temperature. By doing so one obtains:

$$Q = \frac{E_n}{A} =$$

$$\left\{ \rho C_p C_{HE} (U - u_s) \right\} \Delta T_s +$$

$$\left\{\rho\, l_v\, \frac{R_H}{100}\, C_{HE}\, \frac{5487.6}{T_s^2}\, q_s\,(u-u_s)\right\}\Delta T_s + \left\{4\,\epsilon\, T_s^3\right\}\Delta T_s \qquad (3.56)$$

$$\Delta T_s = \frac{E_n}{A}\left\{\rho\, C_{HE}\left[C_p + \frac{R_H\, q_s}{T_s^2} \times \frac{597.6 \times 5487.6}{100}\right](u-u_s) + 4\epsilon T_s^3\right\} \qquad (3.57)$$

where E_n = input of thermal energy and A = lake area.

The equation above may be used to obtain an estimate of the rise in lake temperature for a uniformly-dispersed thermal effluent. The relative magnitudes of the terms give the partitioning of energy between Q_R, Q_{LE}, and Q_H. Thus, for a given undisturbed lake temperature, it is possible to determine the enhanced evaporation knowing the energy input to the lake and the mean wind.

When one considers the case of a plume of uniform surface temperature $T_s + \Delta T_s$ in equilibrium with its surroundings then, neglecting heat-transfer to the rest of the lake, the plume size becomes a function of the wind speed, provided T_s, T_a, R_H, ΔT_s and E_n are known. Once again it is possible to obtain a partitioning of energy between the latent, sensible, and radiative components of energy transfer from the plume surface.

CHAPTER IV

RESULTS

4.1 Introduction

In this chapter the results of the different methods of calculating evaporation from Lake Wabamun are presented and discussed. Before such results can be presented, however, it is necessary to consider the reliability of the measurements. Table 4 gives the estimated errors of measurement of the field instrumentation.

There are major sources of errors other than the intruments. Marciano and Harbeck (1954) found that mid-lake winds were 10% higher than winds measured at the shoreline. By mounting the wind-measuring instruments on berms or protuberances extending into the lake, and then using records from stations for which the wind was off the lake, it was hoped that measurements so obtained would be more representative of mid-lake conditions. Because Lake Wabamun is generally warmer than the overlying air, there is probably a thermally-driven convergent component in the wind field around the lake. Thus, even by taking wind measurements downwind from the lake, there exists the possibility that the winds will be underestimates of the true mid-lake wind. Practical considerations dictated mounting wind-measuring instruments on top of berms or piers

Table 4

Estimates of Measurement Errors

Variable	Estimated Error	Effect of Error on Evaporation Estimate *	Source of Error
Wind speed	0.5 m/s	15 %	Accuracy to which recorder charts could be read
Air temperature	0.5 C	6 %	S.D. of calibration
Relative humidity	4 %	2 %	S.D. of calibration
Water temperature	0.9 C	20 %	S.D. of calibration
Water level	1.3 mm		Reading error

* Figures listed are for conditions which are typical or average for the period of study

extending into the lake. Unfortunately, the physical presence of such a mound on the flat surface of the water would produce an increase in wind speed over the mound, and temperatures and specific humidities measured there would be representative of heights slightly lower than actually measured (Fig 8). An estimate of the greatest error which a ridge the size of the Sundance berm could produce is obtainable from the inviscid-flow approximation for flow perpendicular to a cylinder. From these considerations it is estimated that the winds could be as much as 2% higher over the Sundance berm than those measured at anemometer level away from the berm. The anemometers were calibrated in a wind tunnel, but over the lake the flow was turbulent, and one would expect a degree of overspeeding of the cup anemometers (MacCready 1966). No compensation was made in the calculations or calibration to correct for such overspeeding. The magnitude of such overspeeding would depend on the amount of turbulence present, and hence would be affected to some extent by the stability of the flow. The magnitude of such an overestimate is expected to be less than 10%, however.

The air temperature and humidity as measured by the thermohygrograph mounted in a Stevenson Screen may also be in error due to screen-lag in adjusting to the environment. Bryant (1968) found that the lag time decreased from 30 minutes under calm conditions to 6.5 minutes with a 7 m/s wind. In this study no compensation was made for screen

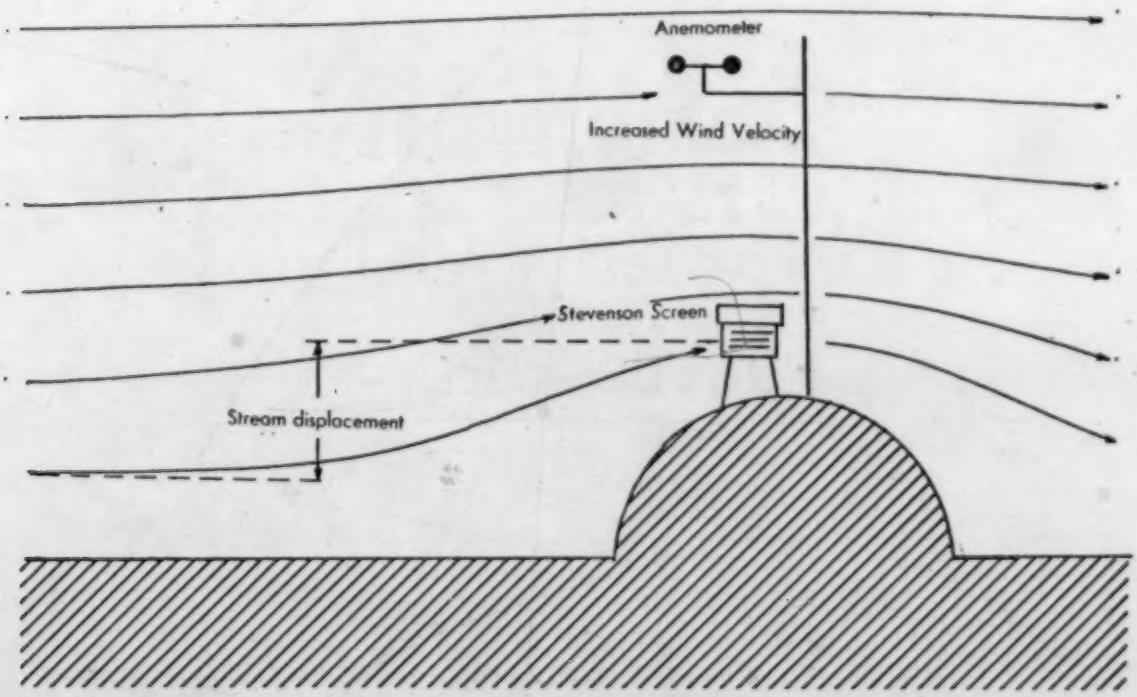


FIGURE 8 Airflow over a mound

lag. Errors due to the neglect of screen lag may have resulted in systematic differences in the calibration of the Sundance and Wabamun thermohygrographs. The weekly calibration measurements were taken shortly before noon at Sundance, and in the mid or late afternoon at Wabamun for most of the checks made of these stations during the summer months in 1974.

Probably the greatest source of error in the water-budget calculations is the question of whether or not the rain-gauge measurements are representative of the lake as a whole. During the summer of 1974 radar observations at the Edmonton International Airport confirmed that shower or thundershower activity occurred on all days for which there were precipitation and radar observations available. It is noted that radar pictures were available only when the radar was working, and when the meteorological technician was not too busy with other duties. For these reasons the radar data were rather incomplete. For water-budget calculations an average of the precipitation measurements from the two stations is used to estimate average lake precipitation. A paired-t test applied to the 78 day total precipitation produced confidence limits of 16.4 ±3.1 cm at the 90% confidence level (16.4 ±6.3 cm at 95% level), assuming that the errors in measurement were randomly distributed. The distribution of rainfall may not be even over the lake due to the interaction of the wind and terrain. There may also be systematic errors in the catch of the tipping-bucket

rain-gauges but the magnitudes of these remain undetermined.

The modified Thornthwaithe procedure (Laycock 1967) based on monthly precipitation and temperature data failed to indicate any runoff from the Lake Wabamun drainage basin. The validity of this procedure when applied to monthly amounts is questionable during two periods of heavy rainfall (day 10 and day 30). When results from water-budget estimates of evaporation are examined there appears to be runoff for up to 10 days after the heaviest rainfall. A period of runoff was suggested by a period of 10 days during which the totaled evaporation decreased. Another possible source of error in water-budget calculations was set-up of the lake surface. A correction was included in lake -level calculations to compensate for set-up of the water surface (Ippen 1966). It was found that the set-up was rather small most of the time. The effect of waves and seiches was not determined, although it is thought that seiches would induce reading errors in the estimation of the mean water level from the chart recording. Waves could induce systematic errors in water level if the Stevens water-level recorders did not respond linearly to variations in water depth.

4.2 Mass-Transfer Estimates of Lake Evaporation

4.2.1 The Bulk-Transfer Coefficients

The bulk-transfer coefficients and evaporation were

calculated from data reduced to hourly values using the theory and procedures outlined in Chapter III. The coefficients, and evaporation were then chosen from either or both stations depending upon whether or not the wind was from the water at the station. The hourly values of the bulk-transfer coefficient at thermograph height are plotted in Fig. 9. A considerable diurnal variation was found for the value of the bulk-transfer coefficient. Weighted daily averages of the bulk-transfer coefficient for evaporation appear in Fig. 10. A comparison of these average coefficients with the daily mean air-water temperature difference (Fig. 11) shows a rough correlation as expected. A mass-transfer weighted average value of the bulk-transfer coefficient for evaporation of 0.00162 was computed for the 78 days of the experiment. This result appears to be in close agreement with the results of Marciano and Harbeck (1954). It must be remembered, however, that the average bulk-transfer coefficient which was calculated is valid at a height of approximately 1.2 m, and that, strictly speaking, the Hefner formula is valid for winds and specific humidities measured at 10 m. An attempt was made to compute an average 10 m bulk-transfer coefficient using the weight function outlined in Chapter III. A weight function for $C_{E_{10}}$ was obtained by taking the ratio of evaporation to the computed 10 m bulk-transfer coefficient. Unfortunately the approximate formula used to calculate the bulk-transfer coefficient produces an unrealistic weight factor for

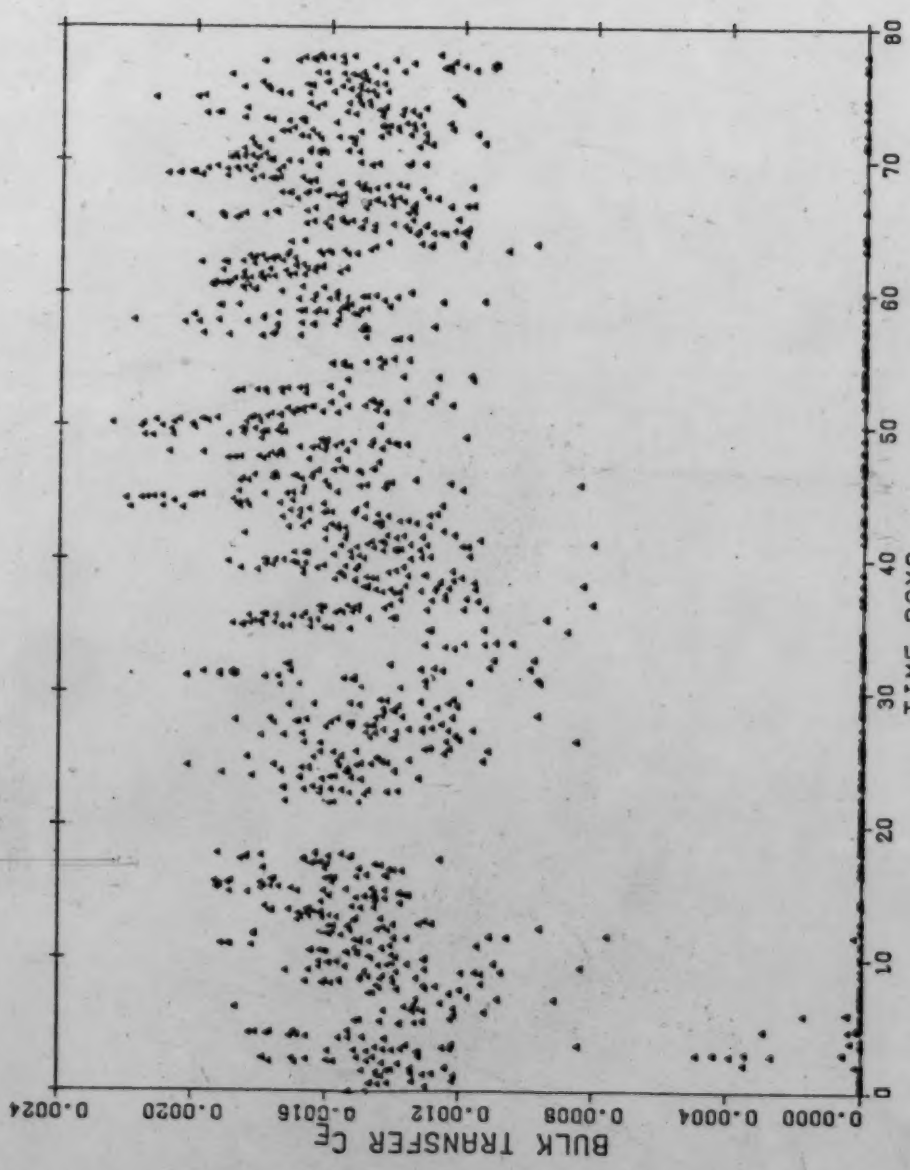


Figure 9. Hourly selected C_E at 1.2 m.

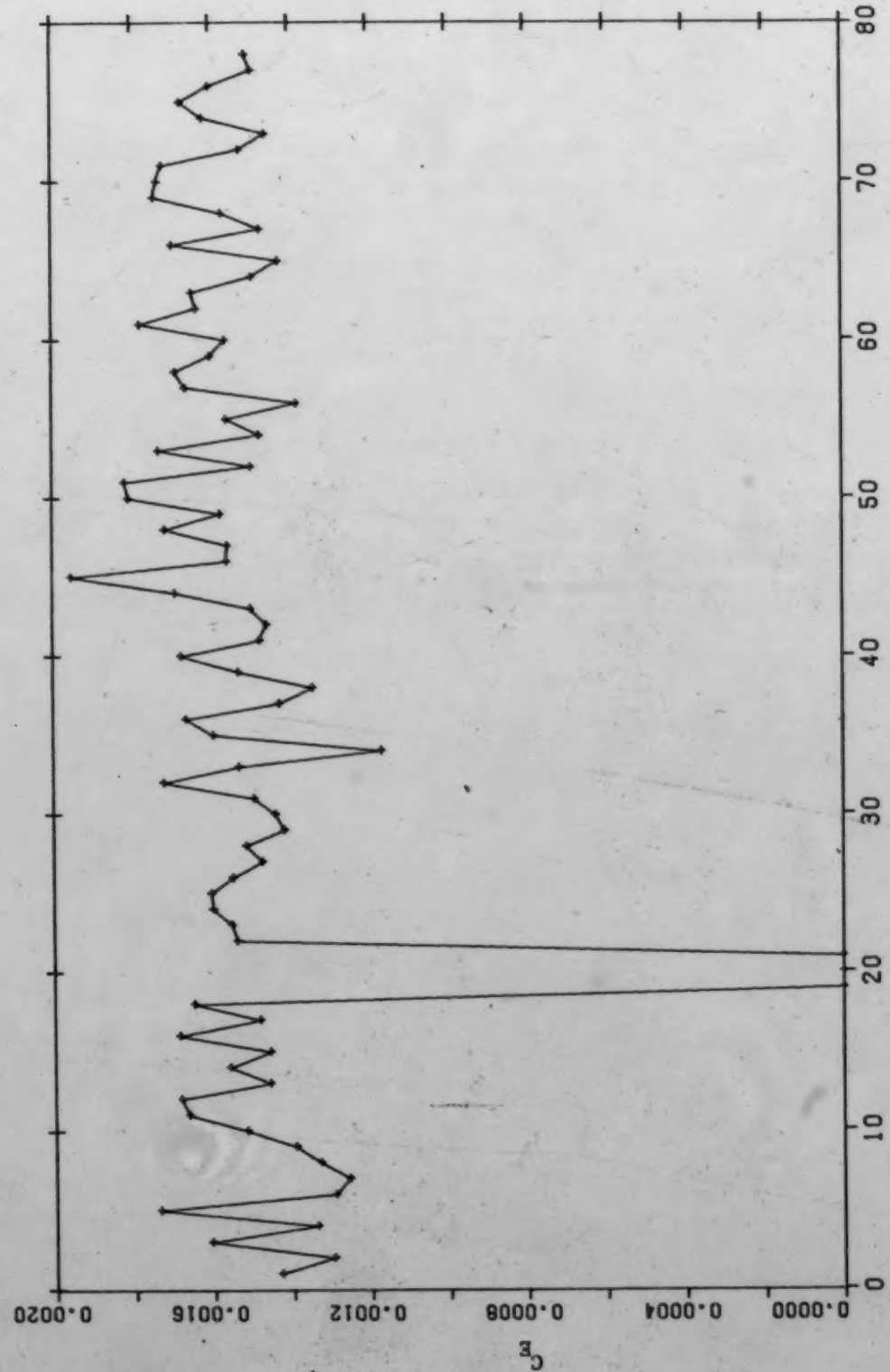


Figure 10. Weighted-average C_E at 1.2 m.

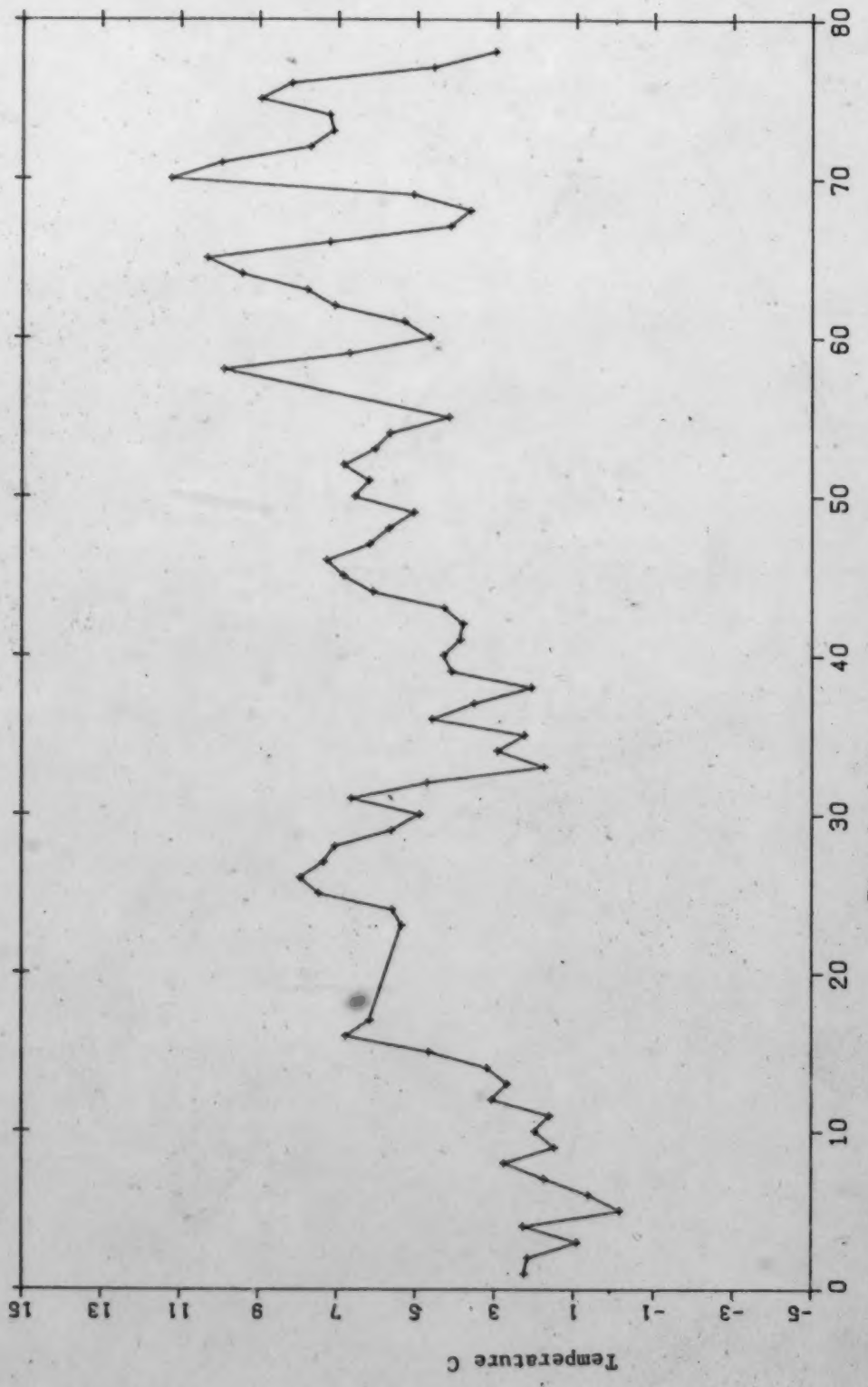


Figure 11. Daily mean air-water temperature difference.

periods of stable stratification. This was because for Ri>1, both evaporation and the bulk-transfer coefficient for evaporation were small. However, the ratio of these two near-zero quantities produces a weight function which is many orders of magnitude larger than weight factors during lapse conditions. It suffices to say that one should not try to divide zero by zero. Weight functions at thermograph height were not plagued by this problem because the wind was interpolated to this height, and the weight function there was obtained by the product of wind and air-water specific humidity difference. The approximation was better at this height because |Ri| was usually less than 1. The evaporation formula that results in a bulk-transfer coefficient of 0.00162 is:

$$E = C_E \rho (u-u_s)(q-q_s) \qquad (4.1)$$

wheras the Hefner formula is:

$$E = \overline{C_E} (u)(q-q_s) \qquad (4.2)$$

Because $u_s = 0.035\ u$ the average bulk-transfer coefficient derived from Lake Wabamun data becomes 0.00156 in terms of the Lake Hefner equation. Most drag-coefficient determinations, including the ones from which the neutral-stability drag coefficient was obtained for this thesis, e.g., Smith and Banke (1975), have obtained the drag coefficient over water without compensating for surface drift. By compensating for surface drift, the neutral-

stability drag coefficients so determined would have been higher by approximately 8%. This and the lack of the surface drift in the Lake Hefner formulation account for about 12% of the observed difference in the evaporation estimated by the two formulas.

Hage (1974) has used 0.0016 as the value for the bulk-transfer coefficient for evaporation to calculate evaporation throughout the open-water season. Such an approach should be justified as long as the stability remains close to that observed in the Lake Hefner experiment at other times in the year. The temperature record shows unstable conditions to exist during the ice-free period and air-water temperature differences like those observed during the period of this study.

4.2.2 The Evaporation

A running total of the evaporation was computed as a change in the lake level using the stability-and-wind-dependent formulation of the bulk-transfer coefficients. The results are plotted in Fig. 12. Because water temperature data were unavailable because of recorder malfunctions, there are two breaks in this record. The evaporation and bulk-transfer coefficients were chosen from either observing station or an average of both, depending upon whether or not the wind was blowing off the lake at the observing site. This was done in an effort to ensure that

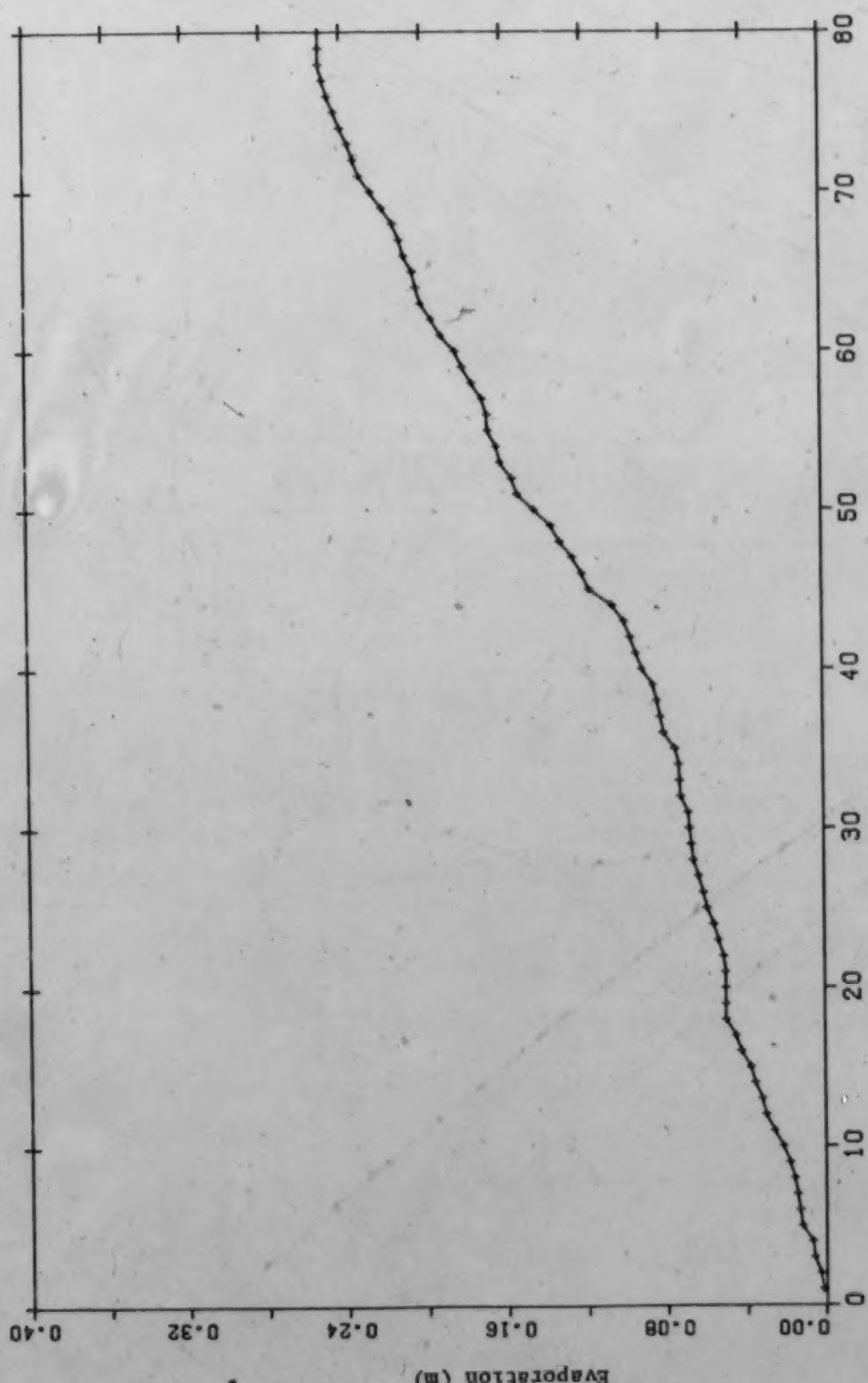


Figure 12. Total Evaporation vs time calculated using the wind-and stability-dependent formula for C_E.

the instruments and measurements were consistently well within the boundary layer of the lake, and thus the conditions for a constant-flux layer would be fulfilled.

Evaporation was also calculated using the same data and the Lake-Hefner equation with a constant bulk-transfer coefficient of 0.0016 (Figs 13,14 and 15). Because wind speed was not measured at 10 m as is required for this fomulation, the wind at 10 m was extrapolated using the wind profile formulation of the previous chapter. For unstable conditions the correction from 3 m to 10 m is small (<2% of u_3). However, for stable conditions the correction can be large. Under stable conditions the wind-profile formulation is not particularly good, because the assumption that z/L be small is not fulfilled. The specific humidity was not corrected to 10 m in this calculation. This would lead to a slight underestimate of the air-water specific humidity difference and the evaporation. Evaporation calculated using the constant bulk-transfer coefficient is plotted as total evaporation in Figs. 13 to 15. Again, gaps in evaporation calculations are due to missing water temperature data. From these graphs of total evaporation it is evident that there is little difference in using either Wabamun or Sundance observing station, or a choice from those stations depending upon whether or not the wind is from the lake.

Some bias is possible in all of the mass-transfer

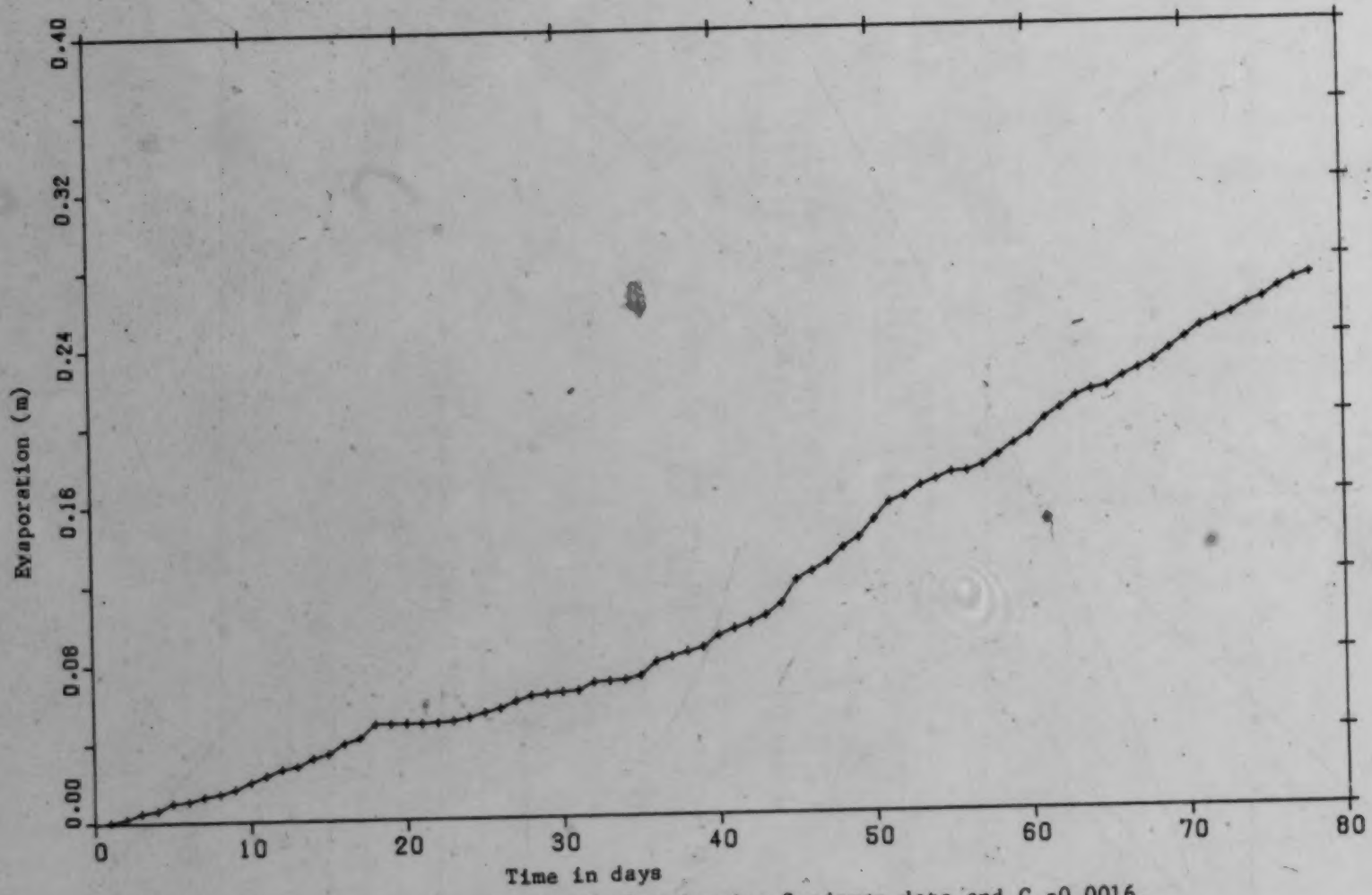


Figure 13. Total evaporation calculated using Sundance data and C_E=0.0016.

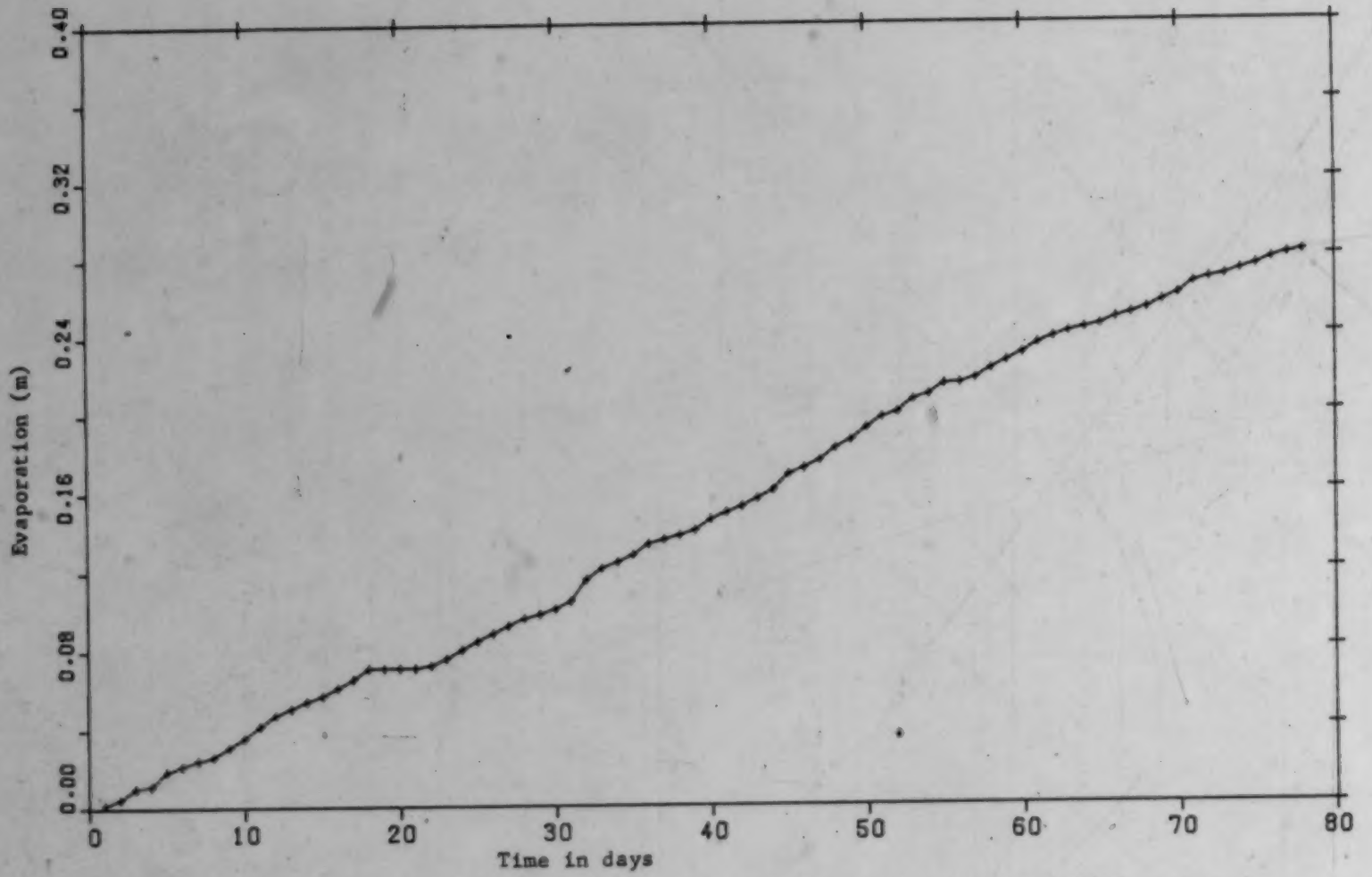


Figure 14. Total evaporation calculated using Wabamun data and C_E= 0.0016.

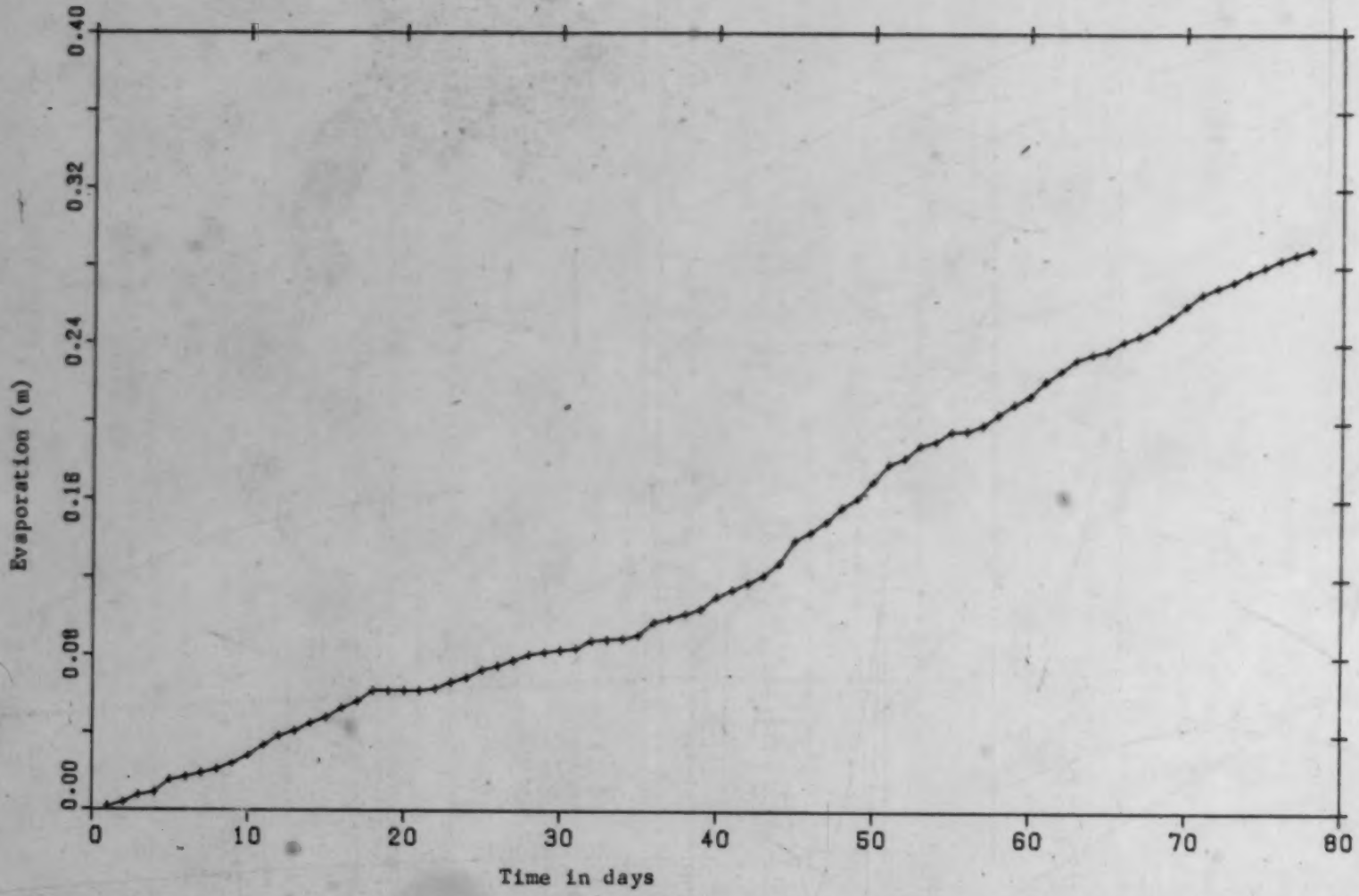


Figure 15. Total evaporation calculated using selected data from Wabamun and Sundance, with C_E=0.0016.

estimates used because of the shoreline location of the thermohygrographs. One would expect that warmer water would tend to remain near the surface, and as a result the warmest water would tend to be dragged by the wind and advected to the lee side of the lake. There was some evidence for the occurrence of this from temperature profile measurements taken by Nuttal (1974). This effect, and the possibility of calibration errors, would tend to account for the differences in evaporation calculated at Sundance and Wabamun. If the wind tends to be onshore at Sundance a greater portion of the time, this station would presumably be within the water-vapour blanket of the lake for a larger portion of time. The specific humidity at Sundance would be higher and the surface specific humidity would be lowered at Wabamun due to upwelling of cooler water. This would partially explain any differences between the two stations.

4.3 The Use of Mean-values of Wind, Temperature and Specific Humidity for Calculating Evaporation.

Rather than use hourly values, it would be more convenient to use daily, weekly or monthly mean winds, humidities and temperatures to compute lake evaporation. The wind-and-stability-dependent formulation is clearly unsuitable for estimating evaporation using mean winds. This formulation, or at least the wind-dependent part of the formulation, is not monotonic with wind speed. Daily mean wind speeds generally would fall on the part of the drag-

coefficient curve (Fig. 7) which is representative of smooth turbulent flow. The wind-and-stability-dependent formulation was not used for estimating evaporation using means of atmospheric variables.

Because $\overline{C_e}$ obtained from the stability-and-wind-dependent formulation approximated the 0.0016 value given by Marciano and Harbeck so closely, this constant value was used to estimate evaporation from Lake Wabamun using mean values of air and water temperatures (Fig 16). It was noted that the use of mean data produces a 10% underestimate of evaporation compared to the use of hourly values. Evaporation was also calculated using the "min-max" method of Spring and Schaefer (1973), but this method produced rather low values (Fig. 17). This method used the Hefner equation, and the daily minimum RH and maximum temperature to compute the daily air-water specific humidity difference and the evaporation.

4.4 Evaporation Estimates Based on the Water Budget Method and a Comparison with Other Estimates

The water budget was used to check the accuracy of the preceding methods of calculating evaporation, and to infer whether or not there may have been a net inflow or outflow of groundwaters to or from Lake Wabamun. To begin with, it was assumed that seepage and runoff were negligible. According to the modified Thornthwaithe procedure of Chapter

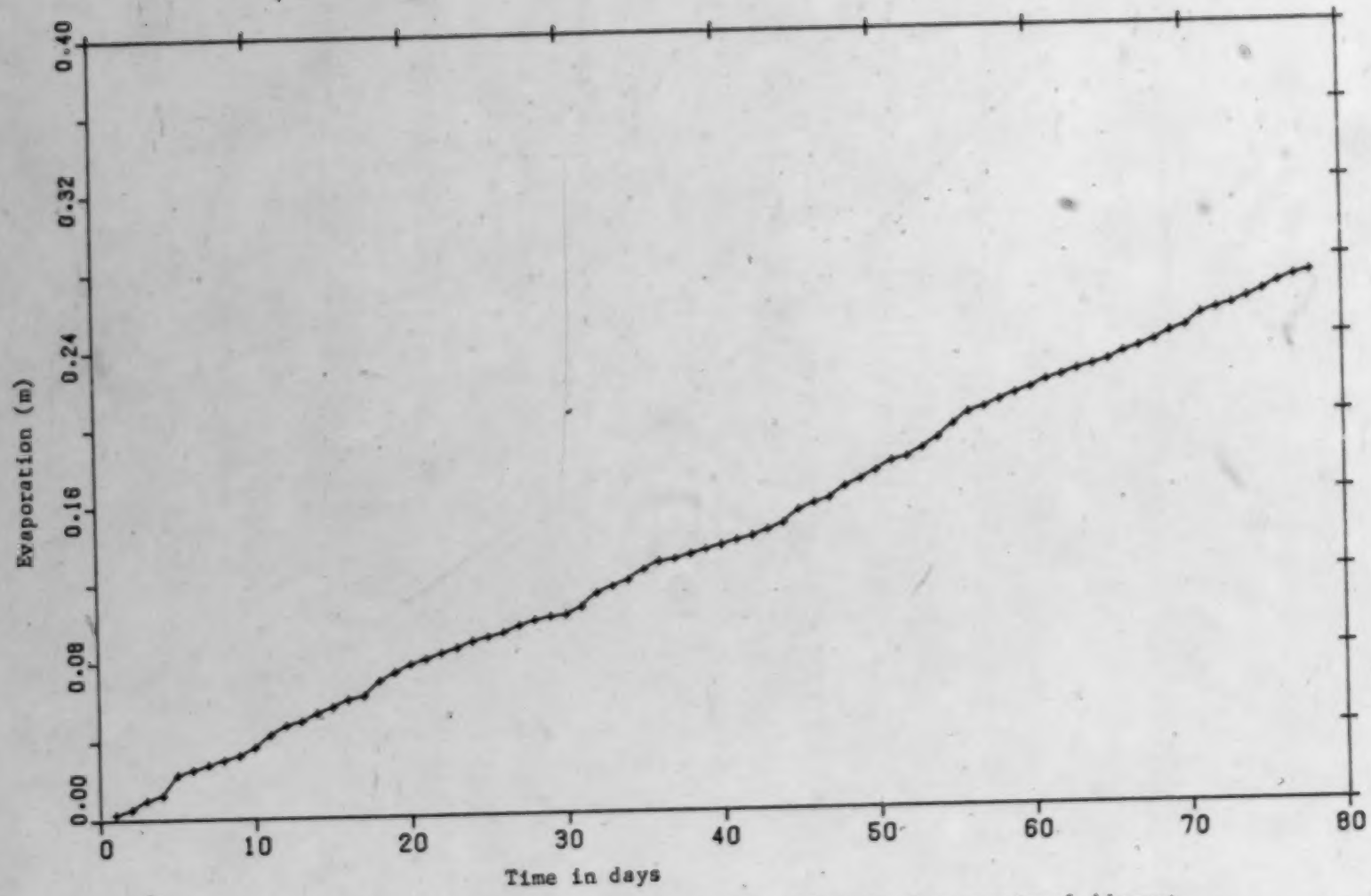


Figure 16. Total evaporation using mean data from Edmonton International Airport.

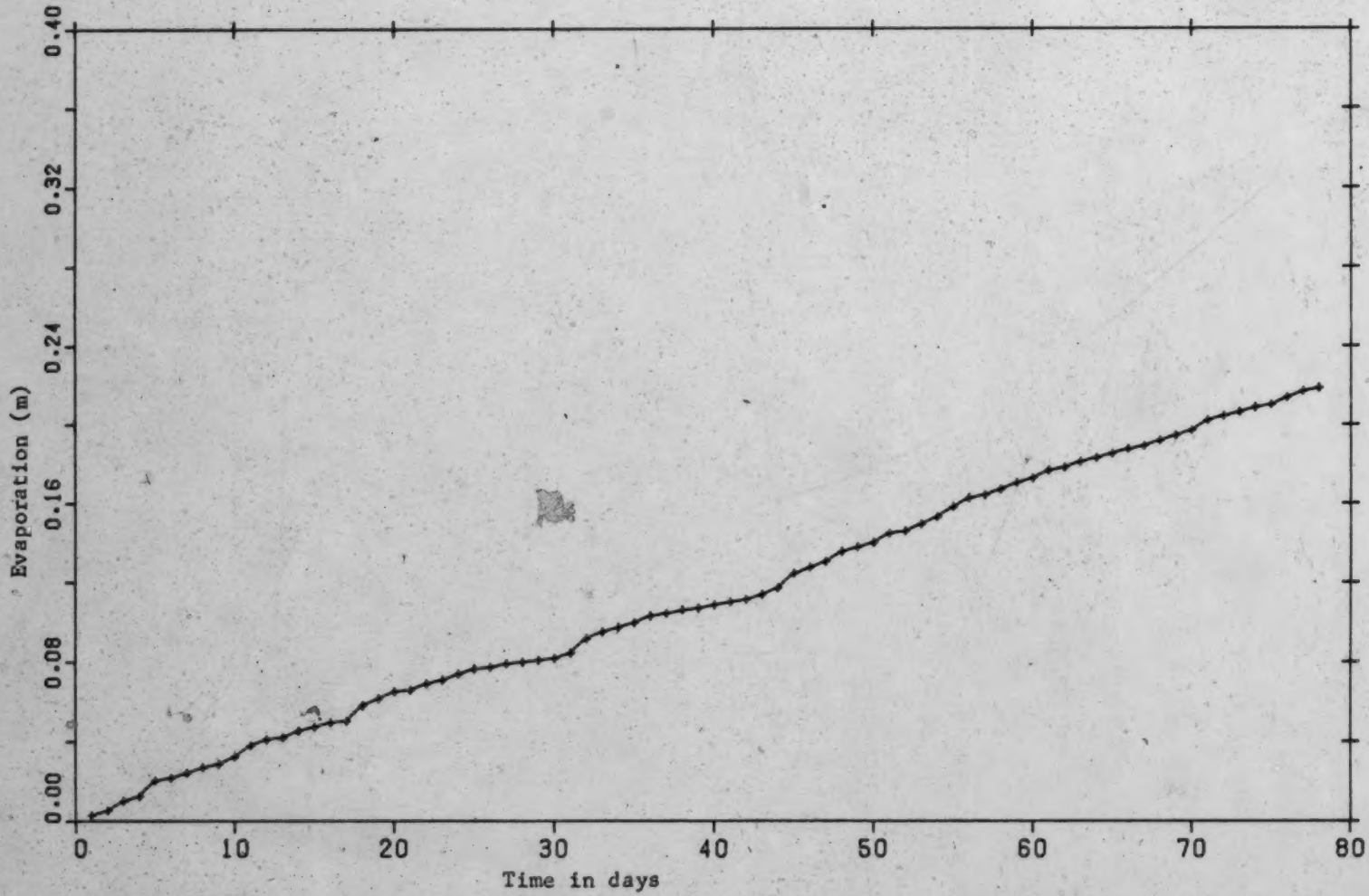


Figure 17. Total evaporation using Edmonton International Airport data and the method of Spring and Schaefer (1973).

III no runoff should have occurred during the course of the experiment. The remaining quantities in the water-budget equation are changes in the water level, the precipitation and the discharge through Wabamun Creek. These quantities are plotted in Figs. 18,19 and 20, respectively. The evaporation is plotted as totalized evaporation because the reading errors of the chart recorder tend to cancel each other on such a plot. It appears that after the major rain on day 30 there was a period of runoff of about 10 days duration. At other times the assumption that runoff was negligible appears reasonable.

The spikes on the plot of total evaporation as a function of time (Fig. 21) are difficult to explain. By comparing the total evaporation (Fig. 21) with the total rainfall (Fig. 19) it appears that these spikes are correlated with rainfall events. A likely explanation is that strong winds associated with these showers produced seiches which made reading of lake level difficult, and introduced noise into the graph of totalized evaporation. This is corroborated by the wind speed observed either prior to, or during the time when such spikes appeared on the water-budget estimate of evaporation.

Using procedures outlined in Chapter III, an estimate for a mean bulk-transfer coefficient was calculated from water-budget evaporation, the air-water specific humidity difference, and the wind speed. Only the periods from day 1

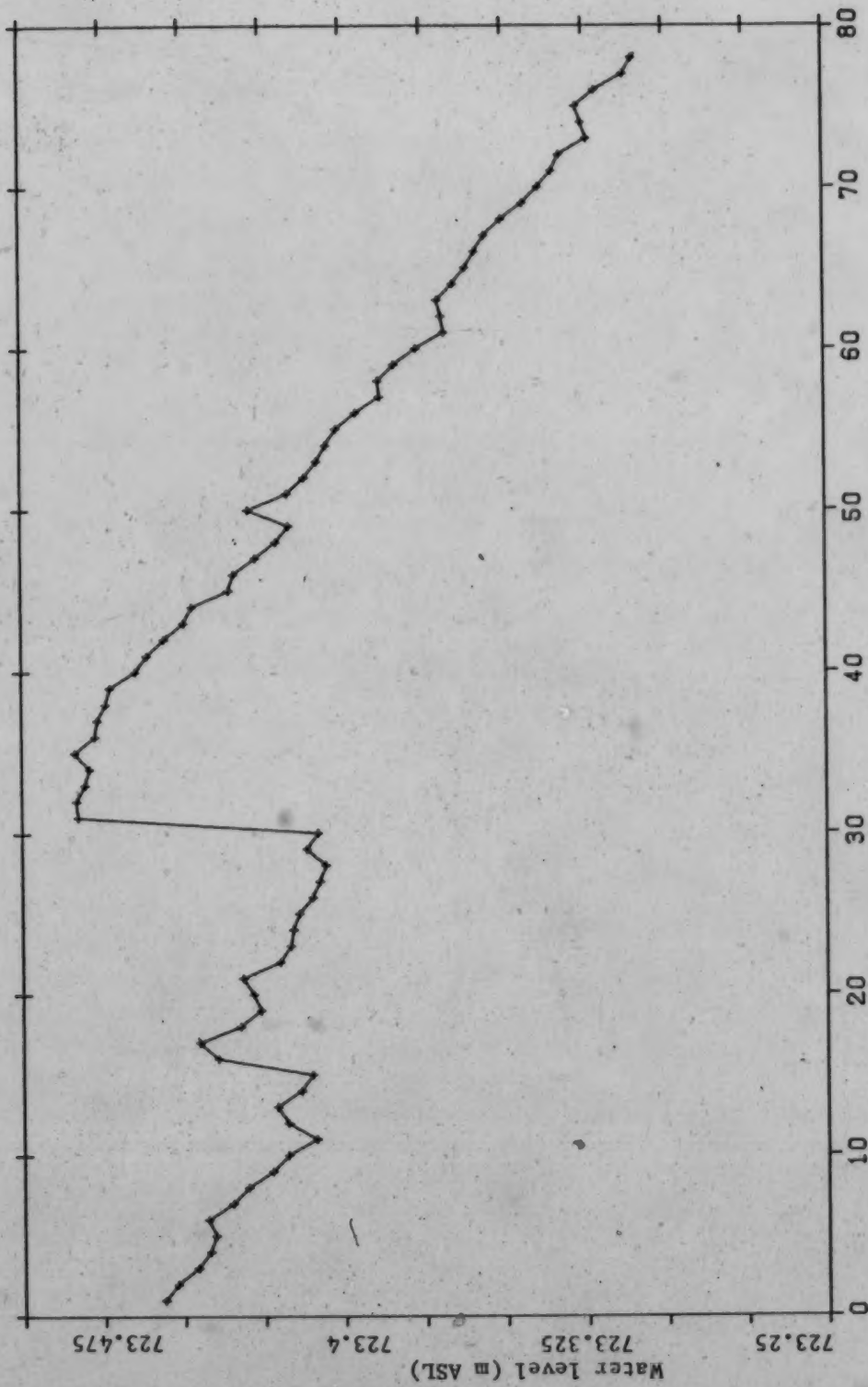


Figure 18. Lake Wabamun water level.

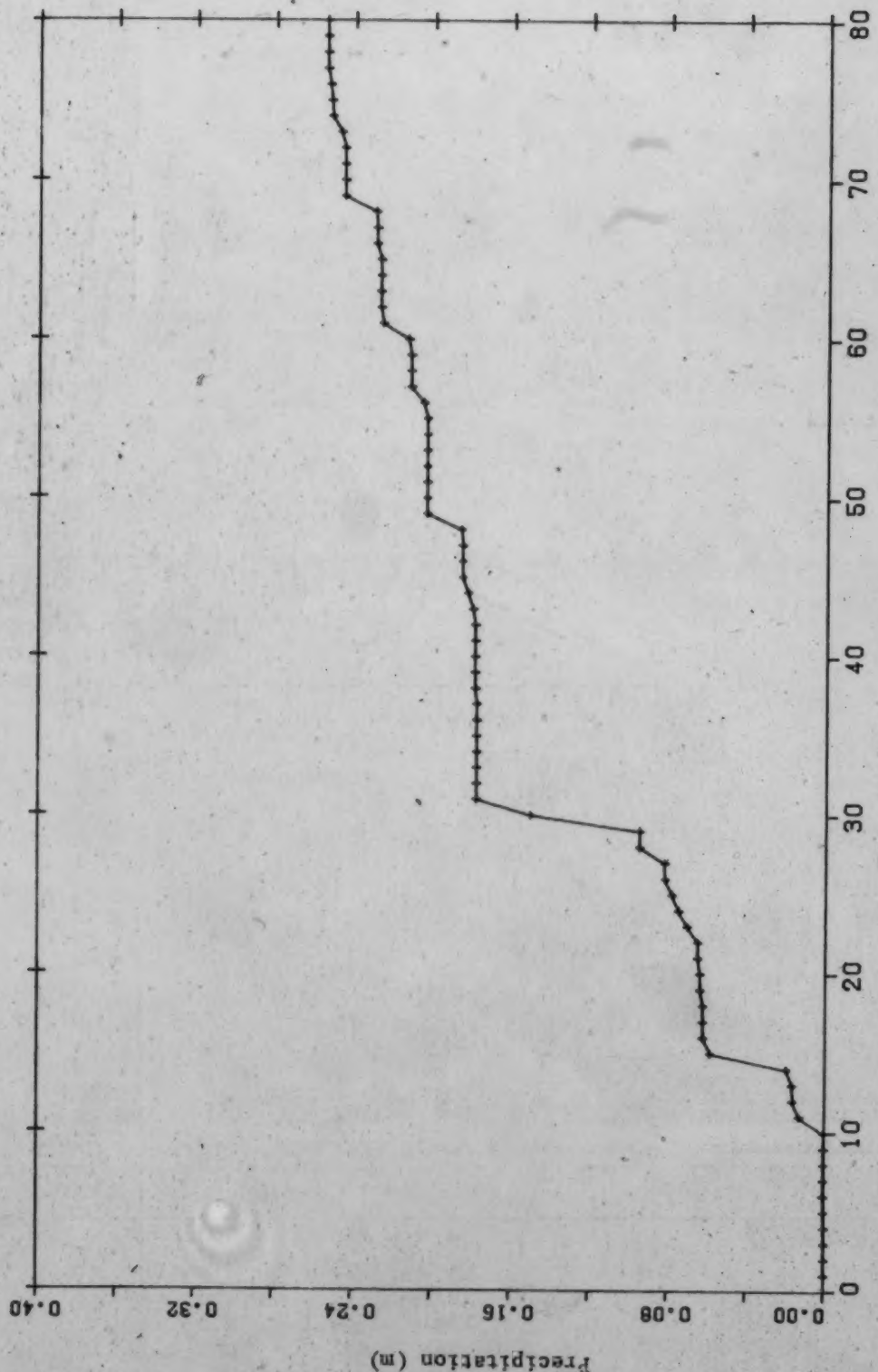


Figure 19. Lake Wabamun total precipitation.

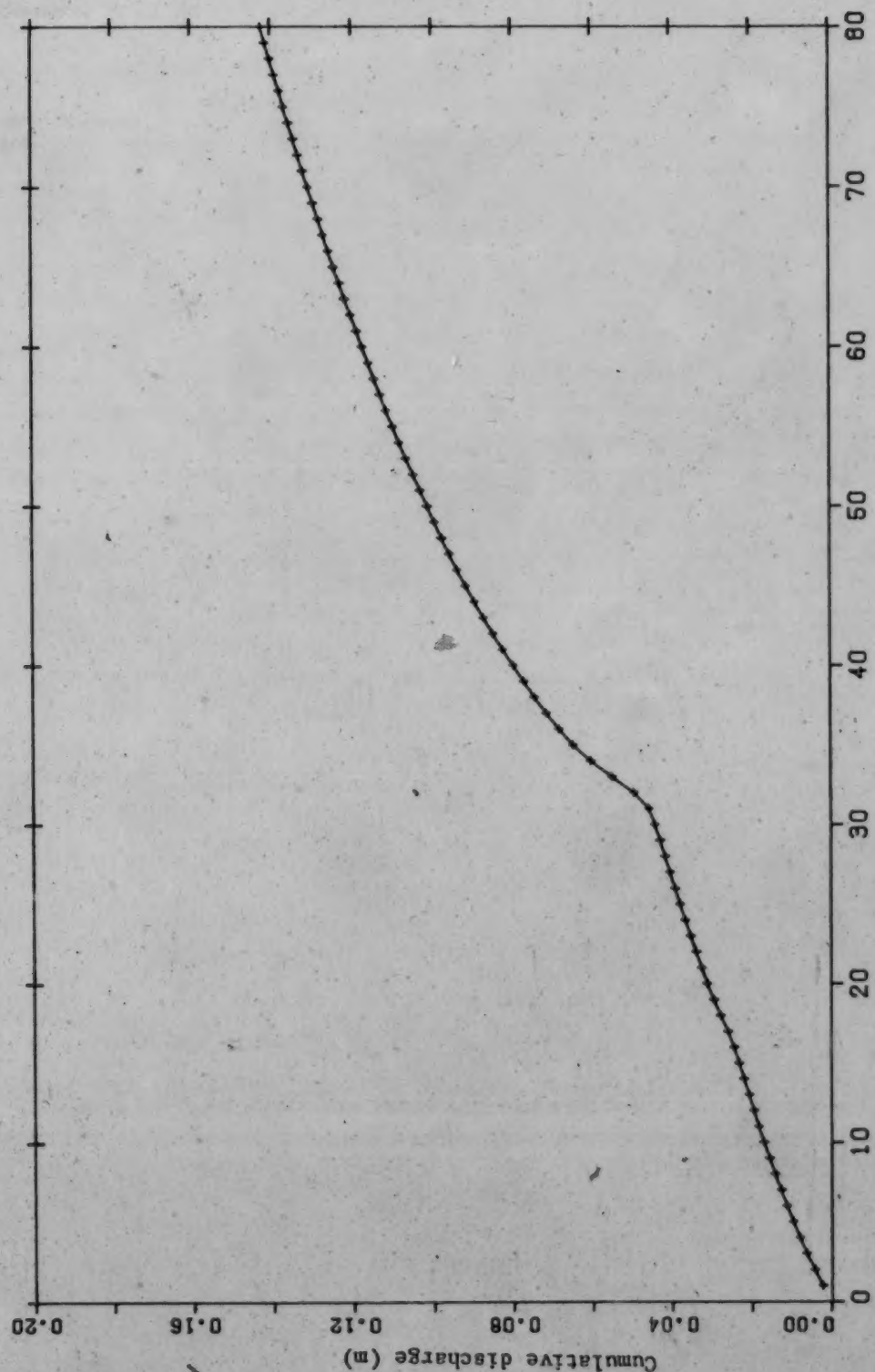


Figure 20. Cumulative discharge through Wabamun Creek.

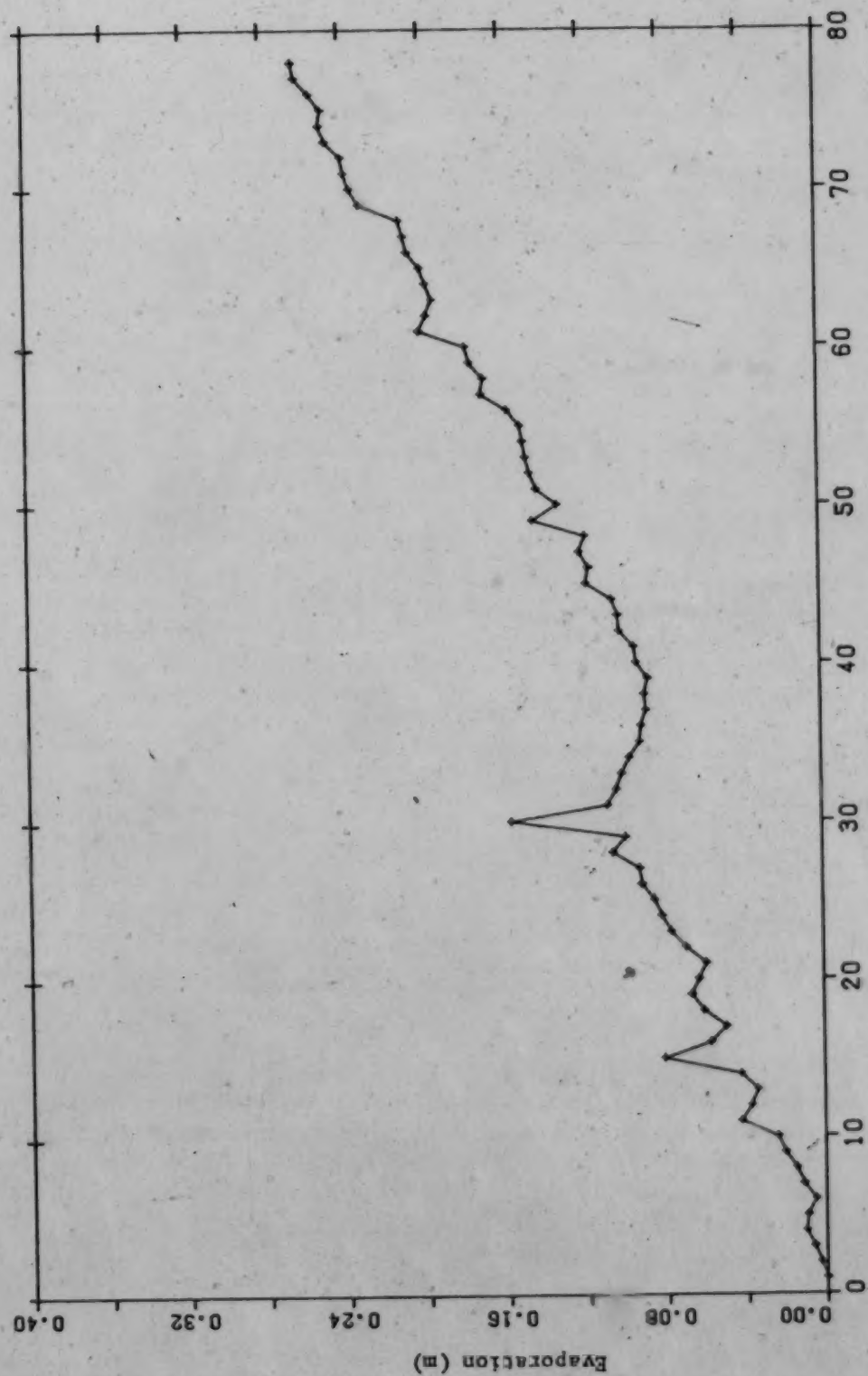


Figure 21. Cumulative evaporation calculated from the water-budget method.

to 17 and from day 40 to 78 were used in the calculations because there was a possibility of runoff from day 30 to day 40 and there was a period of missing water temperature data after day 17. Not much confidence is felt for the early part of the water-budget evaporation estimate, since there was another period where a runoff could possibly have occured around day 10. The bulk-transfer coefficient for evaporation was estimated at 0.0016 ± 0.0006 for days 1 to 17, and 0.0016 ± 0.0001 for days 40 to 78. The error limits on the former figures refer to 90% confidence limits on rainfall obtained from a paired-t test statistic on rainfall at Sundance and Wabamun. In addition to this, the water-budget estimate should have an additional uncertainty of at least 20% since the bulk-transfer coefficient is calculated using the integrated product of wind and air-water specific humidity difference. The evaporation and mean bulk-transfer coefficients computed from the water-budget are about the same as the evaporation and the mean bulk-transfer coefficients of the mass-transfer methods used. Runoff from the small creeks around the lake, if included in the calculations, would serve to make the water-budget evaporation and bulk-transfer coefficients slightly greater.

Considering the confidence limits which the precipitation places on the evaporation, the differences between mass-transfer estimates of evaporation, and water-budget estimates of evaporation, it is not possible to say confidently whether there is a net inflow or outflow of

water from the lake bottom. Previous studies (Nursall, et al., 1972 and Fritz and Krouse, 1973) have also estimated evaporation from Lake Wabamun. Nursall et al. (1972) used monthly means to calculate evaporation from Lake Wabamun and obtained runoff using the method of Bruce and Clark (1966). Bruce and Clark estimate runoff as 4% of precipitation falling on the catchment area of the lake. Hage (1974) found this formula to underestimate snowmelt runoff and to overestimate runoff from showery summer rainfall. A modified Thornthwaite procedure (Laycock, 1967) was used to estimate runoff from Lake Wabamun. The method predicted no runoff, but, as was suggested by Laycock (1975), there could still be runoff associated with a single large rainfall. Laycock also noted that the modified Thornthwaithe procedure could be computed for shorter time periods, and if this were done that it could predict runoff as occurred after the storm rainfall on day 30. However, because there appeared to be a possibility of runoff only after this one event, it was decided to neglect runoff completely in the water-budget calculations, and to concentrate on those times when it was reasonably certain that there was no runoff.

Previous estimates of evaporation from Lake Wabamun vary as to estimates of yearly evaporation amounts, e.g., Nursall, et al. (1972) 51.96 cm, Hage (1974) 54.6 cm, and Perguson (1973) 69 cm. The first two represent evaporation estimates based on monthly averages of atmospheric variables and the bulk-transfer method, while the latter uses the

Meyer formula, and an unknown data base. The preceding results of this chapter indicate that the use of mean values for computing evaporation produces an underestimate of evaporation, when compared to the evaporation computed using hourly values (10% underestimate). This underestimate could be due to the different covariances between winds and temperatures at land and water sites. The evaporation figures of Ferguson appear to be rather high in comparison to the other mass-transfer estimates of evaporation. Hage (1974) used monthly mean winds from the Stony Plain observing site, about 32 km east of the lake. Hage thought this station to be more closely representative of monthly wind speeds at Wabamun. During the course of this study, however, it was found that, over the period of 78 days, the unadjusted mean wind speeds at Wabamun and Sundance were very close to the mean wind speed at the Edmonton International Airport observing site (2.840 m/s, 2.865 m/s and 2.869 m/s, respectively). Hage (1974) used the monthly mean wind speed ratios Edmonton International:Edmonton Stony Plain, and Edmonton Industrial:Edmonton Stony Plain, to adjust Edmonton International and Edmonton Industrial winds to those at Edmonton Stony Plain when the latter were unavailable (prior to 1966). Hage (1974) found that wind speeds at Stony Plain were lower than those at Edmonton, and assumed these would be more representative of the Wabamun area. The results quoted here would tend to question this assumption. However, it must be remembered that this study

covered a period of only 78 days, and that this summer season may not be representative of the climatic normal.

The possibility of a net outflow of groundwater from the lake remains. This possibility was to some extent reinforced by the findings of Nursall, et al, (1972) and Fritz and Krouse, (1973). Because of the isotopic composition of the lake, these authors conclude that the lake is under a strong evaporitic regime. However, estimates for evaporation, precipitation, and runoff computed by these authors require a net inflow of groundwater into the lake. In order to explain the low salinity of Lake Wabamun, these authors postulate that there must also be a discharge of groundwater from the lake. The study of Fritz and Krouse (1973) indicates that wells drilled to the south-west of the lake contain water for which the isotopic composition is similar to the lake water. Based on geological evidence (Carlson, 1970) they conclude that a buried Pleistocene riverbed acts as a recharge and discharge aquifer for Lake Wabamun. They postulate that the aquifer links Lake Wabamun with Lake Isle to the north-west, is supplying Lake Wabamun with water of the right isotopic composition and lower salinity, and is flushing the lake through groundwater to keep the lake at a low level of salinity. The mass-transfer evaporation estimates made in this thesis, when compared to water-budget evaporation estimates, indicate that there is little if any net flow of groundwater either to or from the lake. Furthermore, it is

possible that Lake Wabamun is flushed sufficiently by periodic outflows of water from Wabamun Creek , for which discharge data were incorporated into the water-budget calculations. The creek is intermittent in nature, and discharges water to the North Saskatchewan River whenever the water level on Lake Wabamun is sufficiently high. Hage (1974) found that discharge of lake waters ocurred for lake levels higher than 722.7 m. The findings would indicate that seepage, if it exists, is taking water out of Lake Wabamun and that this seepage, along with periodic discharges of lake water from Wabamun creek, acccunt for the isotopic composition and salinity of the lake. The fact that there is an intermittent discharge from Lake Wabamun indicates that, over longer time periods, the lake receives more precipitation and runoff than is evaporated or seeps through the lake bottom. Assuming a "normal" climate, one could equate this discharge to the allowable evaporation enhancement due to power development.

4.5 Evaporation Enhancement due to Thermal Effluent

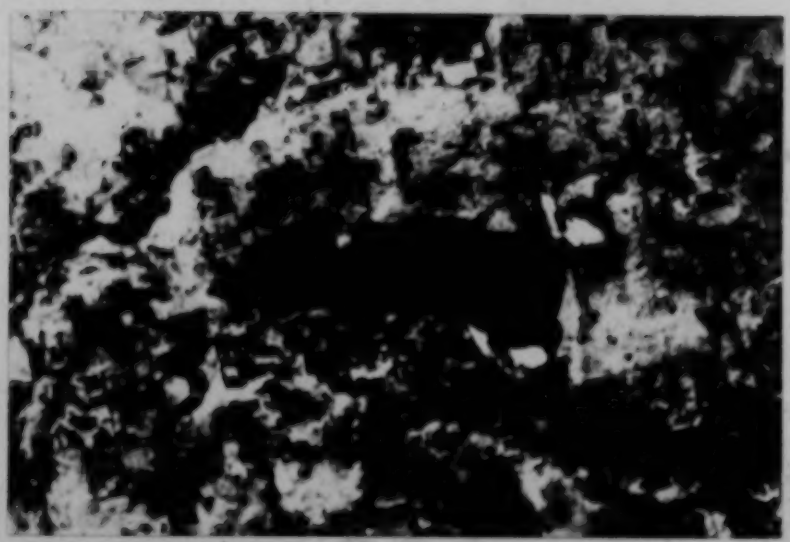
Lake evaporation can be thought of as a mechanism by which the lake gives up energy absorbed as solar radiation. Neglecting the exceptional case where energy is advected to the lake, the incident absorbed solar energy is in equilibrium or near equilibrium with the long-wave terrestrial radiation emission from the lake, and the latent and sensible heat given up to the atmosphere over the long

term. As was formulated in Chapter III any additional energy input to the lake is also partitioned among these components of the energy budget. The power plants on Lake Wabamun put a thermal load of 270 MW into the lake when operating at their maximum capacity of 900 MW. If all this power were to be used exclusively for evaporating water there would be an increase of lake evaporation of 4.1 cm per year. This figure is considerably lower than the 5.8 to 8.1 cm maximum possible yearly evaporation calculated by Ferguson (1973) using the Meyer formula. The high figure may be due in part to an overestimate of wintertime evaporation based on the assumption of too large an open water area by the power plants. This figure may also be affected by Ferguson's interpolation of evaporation data from evaporating-pan stations not representative of the Lake Wabamun area

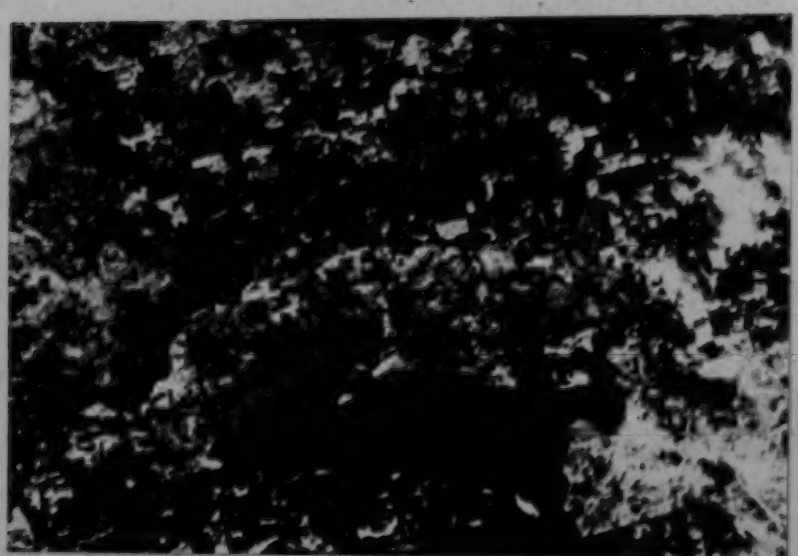
A comment is made at this point that it is conceivable that the open-water area may not be due wholly to the increased heat input of the power plants, but may be a result of the mechanical mixing of the warmer water below the ice by the kinetic energy of the water-jet from the plants to produce an open-water area. This effect would take out heat during the winter months at the expense of energy stored during the open-water season.

Hage (1974) has estimated winter and summer lake evaporation on a monthly basis, and has obtained a yearly

evaporation enhancement of 3.7 cm. Hage used crude estimates of open water and plume areas from the work of Hogg (1973), Nursall and Gallup (1971), and Nuttall (1974). The LANDSAT satellite pictures (Figs. 22 and 23) show the wintertime open-water areas at different times in the winter. Nuttal (1974) has on a few occasions measured water temperatures within the open-water areas. He found that the plume occupied only a small portion of the open-water area. However, these results may not be representative since there was considerable wind on the few occasions that these temperature measurements were made. One would imagine that the stress exerted by the wind on the water would serve to mix the plume and lake water, resulting in a more compact plume. Lal and Rajaratnam (1975b) have determined experimentally the size of a thermal plume in a tank as a function of a plume Richardson Number. The results were of interest, but did not agree very well with the few measurements of the size of the plume taken by Hogg (1973). Not knowing what the appropriate water temperature distribution should be, it was decided to consider two extreme cases of thermal effluent behaviour as outlined in Chapter III, and to compute the relative magnitudes of enhanced energy transfer by radiation and latent and sensible heat transfers. Because the energy input is known, this analysis should set upper and lower limits on evaporation enhancement. Figs. 24 to 26 show the relative percentages of the energy transfer components, and the rise



Date: 30 Dec. 1973



Date: 17 Jan. 1974

FIGURE 22. LANDSAT satellite pictures of Lake Wabamun.

Date: 21 Feb. 1974

Date: 17 Apr. 1976

FIGURE 23. LANDSAT satellite pictures of Lake Wabamun.

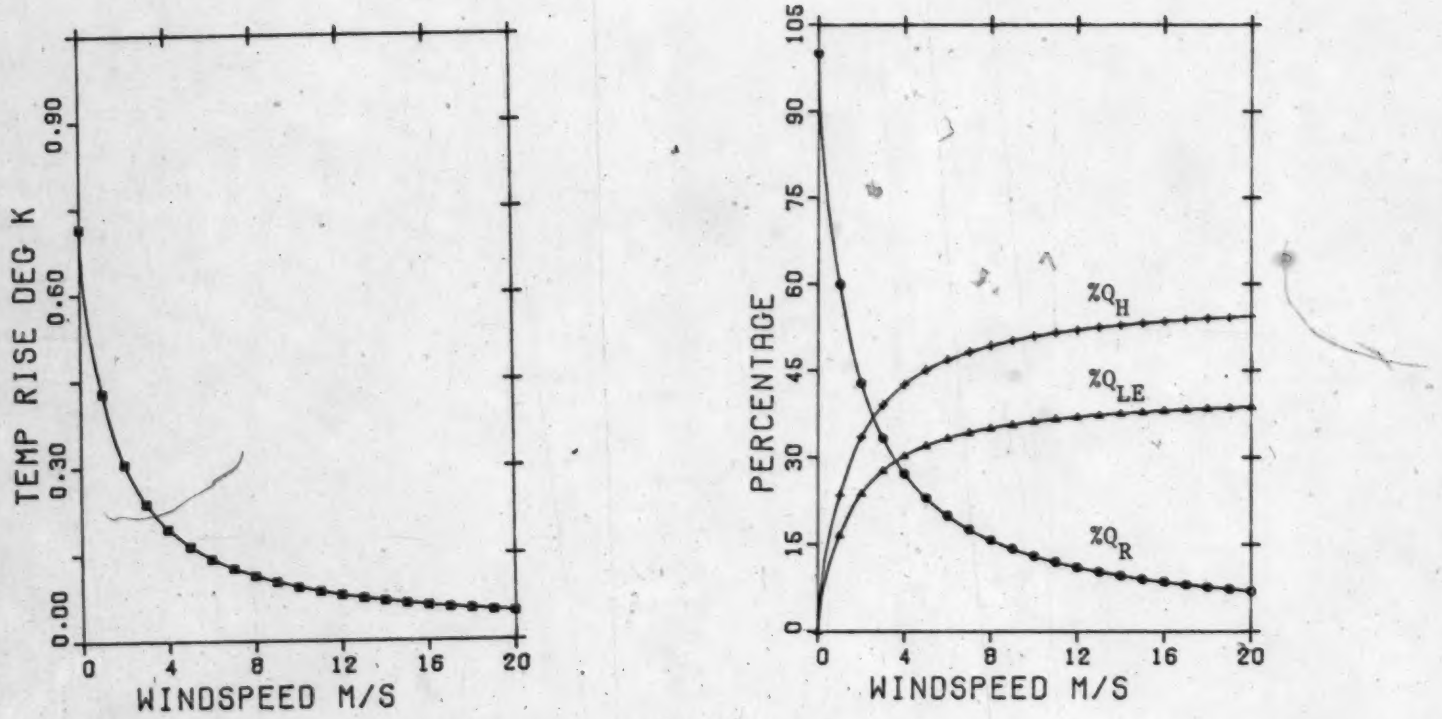


FIGURE 24. Lake equilibrium temperature rise and energy transfer percentages T_s =0 C.

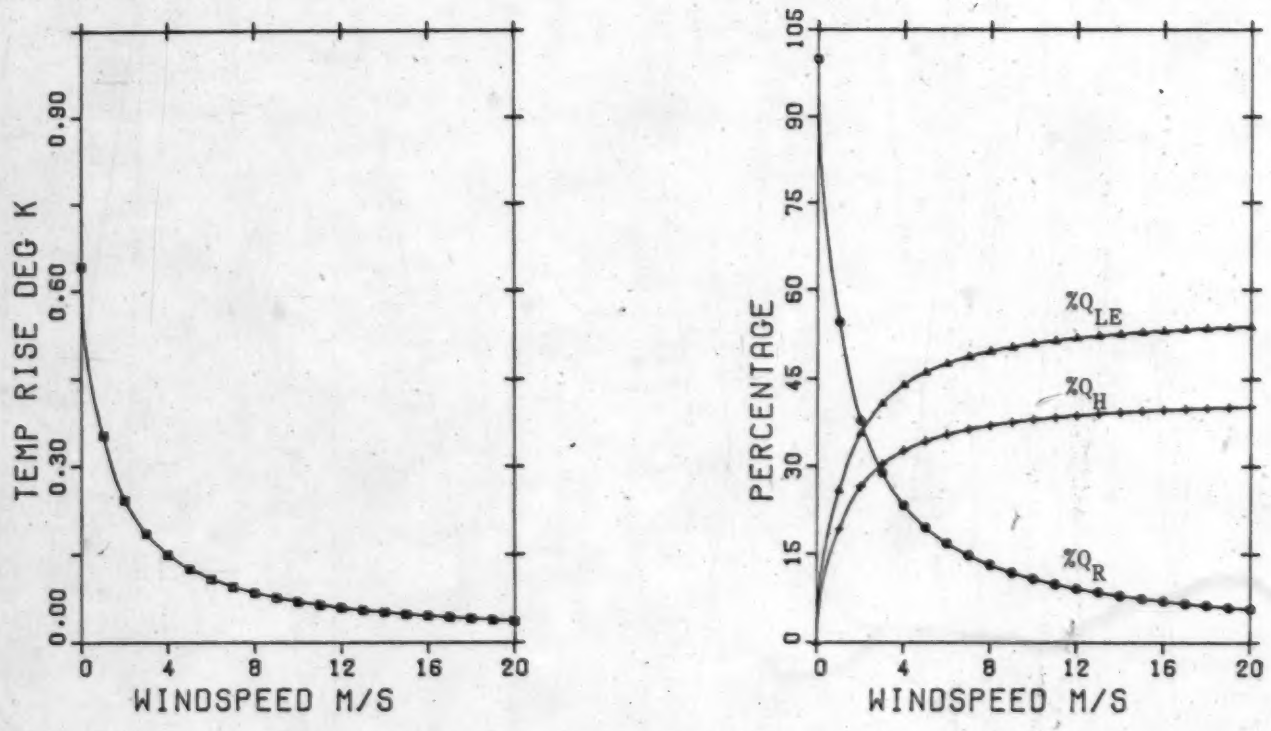


FIGURE 25. Lake equilibrium temperature rise and energy transfer percentages T_s = 10 C.

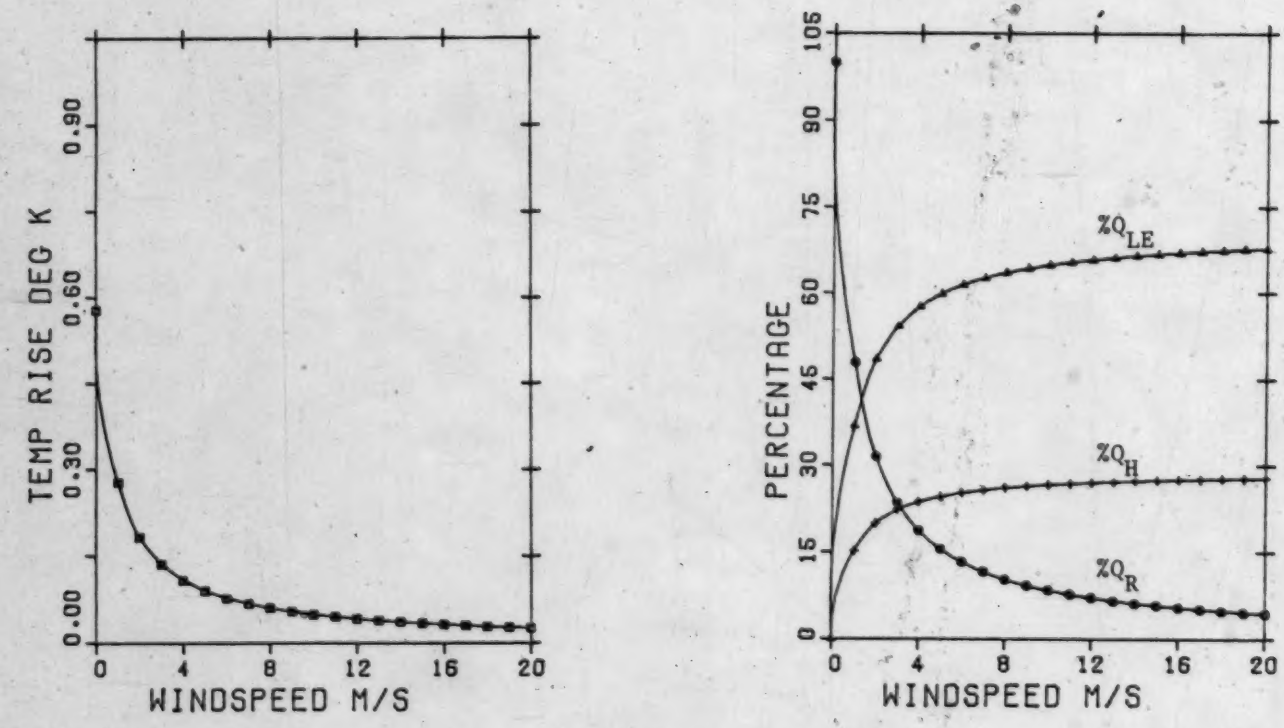


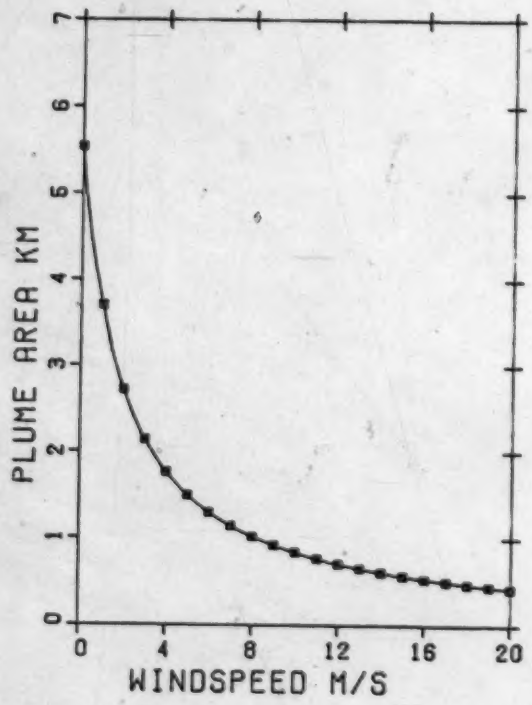
FIGURE 26. Lake equilibrium temperature rise and energy-transfer percentages T_s = 20 C.

in temperature of Lake Wabamun as a function of wind speed. If we look at a typical wind speed of 4 m/s, the lake temperature is seen to rise by 0.18 C for an unpolluted surface temperature of 0 C, and .12 C in summer for a water-surface temperature of 20 C. The percentage of extra heat input dissipated by latent heat rises from 30% in winter to 58% in summer. A crude, all-season average gives a figure of 44% for the portion of total energy input by power plants that is dissipated as latent heat.

Figs. 27 to 29 show the relative percentages of the energy transfers, and the size of the thermal plume required to dissipate the excess energy of the power plants, assuming a plume temperature 10 C higher than the water temperature. In this case, latent heat accounts for 35% of energy loss in winter and 60% in summer. A crude, all-season average gives 48% as the percentage of energy input that is dissipated as latent heat.

From the results above one would expect that latent heat accounts for slightly less than 50% of the thermal pollution from the power plants, and that enhanced evaporation should be no more than about 2 cm/year.

Because Lake Wabamun occasionally discharges water from Wabamun Creek during years with high water levels, it should be possible to estimate an average discharge of water from the lake. If the normal discharge is greater than that required to alter the lake level by 2 cm/year, then there



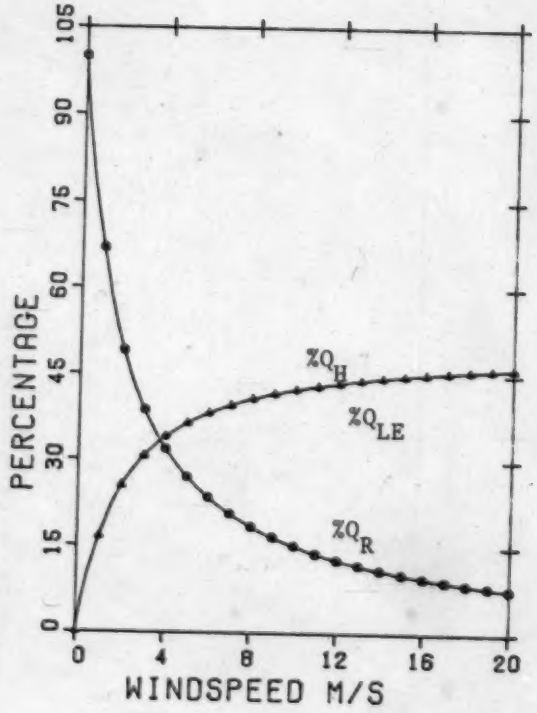


FIGURE 27. Plume area and energy transfer percentages T_s=0 C.

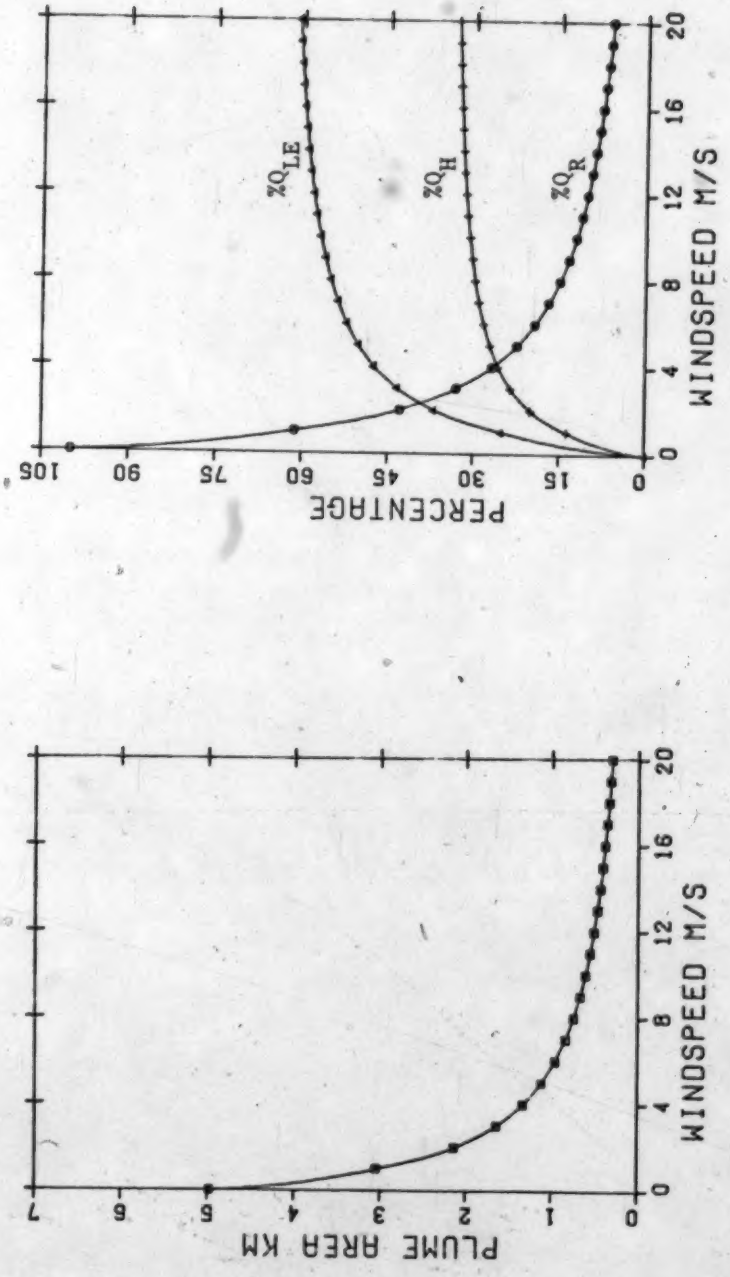
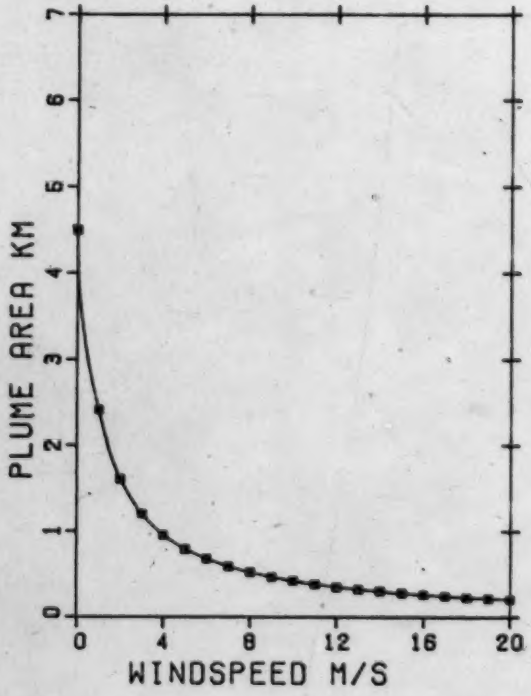


FIGURE 28. Plume area and energy transfer percentages $T_s = 10$ C.



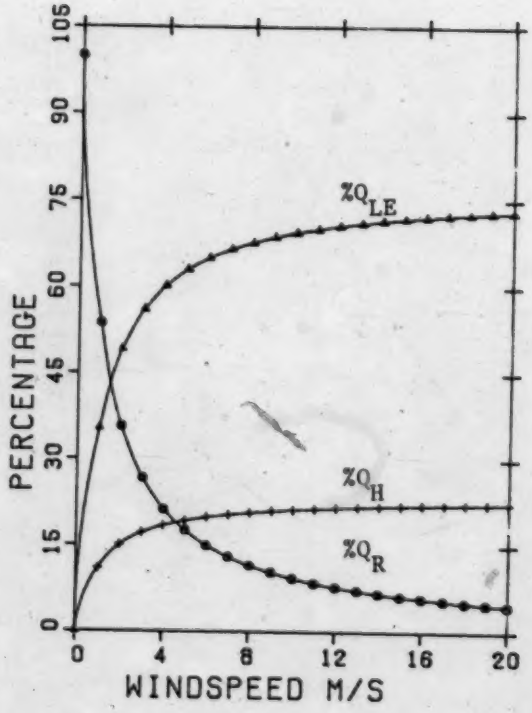


FIGURE 29. Plume area and energy transfer percentages T_s =20 C.

should be little or no problem of enhanced evaporation associated with the operation of the power plants at their present capacity.

CHAPTER V

CONCLUSIONS

The period of time covered by this study was short, and restricted to the summer months. However, it is doubtful that a longer period of study would yield a refinement of the conclusion that the wind-and-stability-dependent version of the bulk-transfer coefficient does not yield a better evaporation estimate than a constant bulk-transfer coefficient. Within the uncertainties imposed by measured components of the water budget, either estimate produces equally valid results.

It was encouraging to find that the time-averaged value for the bulk-transfer coefficient determined from profile theory was close to the 0.0016 value from the Lake Hefner studies which was used by Hage (1974) in his estimation of Lake Wabamun evaporation. It is also concluded that the use of daily averages of temperature and humidity data from Edmonton International Airport was justified to within the accuracy of this study. While these results were encouraging, it is necessary to point out that much of the theory used to construct the wind-and-stability-dependent bulk-transfer coefficient requires some validation and refinement. In particular, the scaling lengths z_o and z_t, and their interdependence, need to be known explicitly over

a water surface. The stability dependence of K_h/K_m needs to be determined in a more accurate and unambiguous manner.

A limit for enhanced evaporation due to the injection of heated waters was calculated. It was found that a little less than 50% of the thermal load on the lake was used to evaporate water, The remaining 50% is dissipated as sensible heat and terrestrial radiation. With a power-generating capacity of 900 MW, lake evaporation is increased by at most 2 cm yearly. A mean discharge of 6 cm annually was computed for Wabamun Creek. Taking this discharge to be surplus to that required to maintain the present lake level, it should be possible to double or triple the present power generating capacity without having to import water to maintain lake levels. Lake-level decreases would be greater during dry years. However, the decreases could be at most double the present lake-level decreases. Perhaps, for this reason, and because of the impacts of the hot plume on the local plant and animal life, the added expenses incurred by Calgary power in diverting North Saskatchewan River water to cooling ponds is reascnable and justified.

LIST OF SYMBOLS

SYMBOL	DESCRIPTION
A	area
C, C_D, C_H, C_E	bulk-transfer coefficient, drag coefficient, and bulk-transfer coefficients for heat and water vapour, respectively
$(C_{ }_{ })_N$	transfer coefficient under neutral conditions
C_p	specific heat of air at constant pressure
E	evaporation
E_p	energy input to thermal plume
g	acceleration due to gravity
I_p, I_R, I_G	precipitation, runoff and ground water inflow expressed in terms of lake-level change, respectively
K_M, K_H, K_E	eddy diffusivity for momentum, heat and water vapour, respectively
k	von Karman's constant
L	Monin-Obukhov Length
O_E, O_G, O_A	evaporation from the lake surface, subsurface outflow and streamflow from the lake, respectively
p	atmospheric pressure
Q, Q_R, Q_{LE}, Q_H	total heat flow, heat flow due to radiative, latent and sensible heat transfers, respectively
Q_{RS}, Q_{RL}	solar and terrestrial radiative transfers
q, q_s, q', q_z	specific humidity, saturation specific humidity of a water surface, fluctuating and at a level z, respecively

SYMBOL	DESCRIPTION
R_e, R_{e_0}	Reynolds and interfacial Reynolds numbers, respectively
$(Ri)_B$	Bulk Richardson number
RH	relative humidity
t	time
$T, T_s, T', T_v, T_v', T_z$	temperature, surface, fluctuating, virtual, fluctuating virtual and temperature at a level z, respectively
u_s, u_z	wind at the surface and at a level z, respectively
u, v, w, u', v', w'	horizontal and vertical wind components and fluctuations of these, respectively
u_*	friction velocity
z	height above ground (m)
z_o, z_q, z_T	roughness, or scaling lengths for momentum, moisture and heat, respectively
ϵ	emisivity
ΔL_z	change in surface elevation of the lake
ν	kinematic viscosity
ρ	air density
σ	Stefan-Boltzman constant
τ, τ_o	shear stress and surface stress, respectively
ϕ_1, ϕ_2, ϕ_3	dimensionless gradients of wind, temperature and water vapour, respectively

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APPENDIX

SUBROUTINE TO SOLVE FOR THE

BULK-TRANSFER COEFFICIENTS

List of FORTRAN Variables

Variable	Description
TTH	the height above water of the themohygrograph
QMQS	air-water specific humidity difference
TV	virtual temperature
UAN	wind speed at anemometer height
ZTH	height of the thermohygrograph above water
ANHYDF	difference in heights between the anemometer and thermohygrograph
CEOVCH	the ratio C_E/C_H (=1)
CDN	the neutral-stability drag coefficient
RI	the bulk-Richardson number
CD	the drag coefficient
CE	the bulk-transfer coefficient for evaporation
UTH	wind speed at thermohygrograph height relative to the water surface
TW	the water surface temperature
EVAP	the evaporation rate expressed as a lake level change in m/s
U10	the wind speed at 10 m
CE10	the bulk-transfer coefficient for evaporation at 10 m
OBML	the Monin-Obukhov length
CDN10	the neutral-stability 10 m drag coefficient
CD10	the 10 m drag coefficient

```
C         THIS FILE CONTAINS THE SUBROUTINE WHICH FITS A PROFILE, AND
C         FINDS THE BULK TRANSFER COEFFICIENTS SUITIBLE TO THE OBSERVED
C         DATA.
C
            SUBROUTINE FUDGE
          DIMENSION D(7)
         COMMON TTH(24,150),QMQS(24,150),TV(24,150),UAN(24,150),ZTH(24,150),
        1 ,ANHYDF,CEOVCH,CDN(24,150),RI(24,150),GU,GS,
        2  CD(24,150),CE(24,150),UTH(24,150),TW(24,150)
        3 ,IB,IDAYS,EVAP(24,150),U10(24,150),CE10(24,150),OBML(24,150),
        3 CDN10(24,150),CD10(24,150),CH(24,150),HEAT(24,150),CH10(24,150)
           REAL L
 700   FORMAT(' DAY ',I3,' IS BEING COMPUTED')
         DO 102 J=IB,IDAYS
         WRITE(7,700)J
         DO 102 I=1,24
C
C        CHECK FOR BAD DATA
C
         IF(TTH(I,J).LE.0.0.OR.TV(I,J).LE.0.0.OR.UAN(I,J).LE.0.0)GO TO 110
          IF(ZTH(I,J).LE.0.0 .OR.TW(I,J).LE.0.0)GO TO 110
C
C         BEGIN COMPUTING THE BULK-RICHARDSON NUMBER. THE SIGN OF RI AND
C         THE QUANTITY BRAC ARE THE SAME.
C
           BRAC=(9.8/TV(I,J))*(TTH(I,J)-TW(I,J)+0.61*TTH(I,J)*CEOVCH*
        1  QMQS(I,J))
```

```
C
C       BELOW AN ESIMATE FOR USTAR IS COMPUTED BASED ON THE LOG-LAW.
C       FIRST GUESS FOR THE SHEAR VELOCITY
C
        USTAR=.035*UAN(I,J)
          N=0
        ZAN=ZTH(I,J)+ANHYDF
 100        U=UAN(I,J)-USTAR
         N=N+1
        CDA=FCDN(ZAN,U)
         USTARO=USTAR
        USTAR=U*((CDA)**.5)
         IF(N.GT.20)GO TO 501
        IF(ABS(USTAR-USTARO).GT.(.005*USTAR))GO TO 100
 501    CONTINUE
         U=UAN(I,J)-USTAR
        CDA=FCDN(ZAN,U)
         UH=(USTAR/.4)*ALOG(ZTH(I,J)/ZAN)+U
C
C       FIRST ESTIMATE FOR Z/L
C
        ZOVL=0.4*((CDA**(-.5))*BRAC*ZTH(I,J))/(UH*UH)
        L=ZTH(I,J)/ZOVL
          M=0
         UANT=0.0
 104       CONTINUE
         USTARO=USTAR
```

```
          M=M+1
C
C       DEPENDING OPON WHETHER IN THE STABLE, OR UNSTABLE REGION
C       USE THE APPROPRIATE PROFILE FORMULATION.
C
          IF(L.LT.0.0)GO TO 200
           GO TO 203
  200       CONTINUE
          H=+ANHYDP/6.0
           DO 201 NN=1,7
           ZED=ZTH(I,J)+FLOAT(NN-1)*H
           D(NN)=USTAR/((.4*ZED)*(1.0-GU*ZED/L)**.25)
  201       CONTINUE
C
C       A NUMERICAL INTEGRATION FOR THE UNSTABLE CASE
C
            UANT=UH+.3*H*(D(1)+D(3)+D(5)+D(7)+5.0*(D(2)+D(6))
     1 +6.0*D(4))
           GO TO 204
C
C       THE LOG-LINEAR EQUATION FOR THE STABLE CASE
C
  203   UANT=UH+(ALOG(ZAN/ZTH(I,J))+GS*ANHYDP/L)*USTAR/.4
  204      CONTINUE
        UH=(U-UANT)*ZTH(I,J)/ZAN+UH
        UHG=UH+USTAR
C
```

```
C       ITTERATE TO OBTAIN A USTAR TO FIT THE PROFILE
C
        DO 140 IN=1,20
        CDNH=FCDN(ZTH(I,J),UH)
        RIB=BRAC*ZTH(I,J)/(UH*UH)
        CDH=FCD(CDNH,RIB)
       US=USTAR
       USTAR=UH*(CDH)**.5
        UH=UHG-USTAR
       IF(ABS(USTAR-US).LT..002*(USTAR))GO TO 141
 140         CONTINUE
 141    CONTINUE
        ZOVL=.4*(CDNH**(-.5))*RIB
        L=ZTH(I,J)/ZOVL
         U=UAN(I,J)-USTAR
        IF(M.GT.20)GO TO 105
C
C       IF THE PROFILE HAS BEEN FITTED SUFFICIENTLY ACCURATELY
C       CEASE ITTERATION
C
        IF(ABS(UANT-U).LT.(.005*UANT))GO TO 105
        UANT=U
        GO TO 104
 105      CONTINUE
C         THE NUMERICAL SOLUTION OF THE ANALITICALLY INSOLUBLE EQUATIONS
C       DETERMINING UTH ETC. IS NOW COMPLETE
C
```

```
          UTH(I,J)=UH
          CDN(I,J)=CDNH
          RI(I,J)=RIB
          CD(I,J)=CDH
          CE(I,J)=FCE(CDNH,RIB)
301       CONTINUE
C
C     COMPUTE 10M DRAG COEFFICIENTS, AND WINDSPEEDS.
C
          CH(I,J)=FCH(CDNH,RIB)
          IF(L.GE.0.0)GO TO 300
          H=(10.0-ZAN)/6.0
          DO 301 K=1,7
          ZED=ZAN+FLOAT(K-1)*H
          D(K)=USTAR/((.4*ZED)*(1.0-GU*ZED/L)**.25)
          U10(I,J)=U+.3*H*(D(1)+D(3)+D(5)+D(7)+
         1 5.0*(D(2)+D(6))+6.0*D(4))
          GO TO 302
300       U10(I,J)=U+(ALOG(10.0/ZAN)+GS*(10.0-ZAN)/L)*USTAR/.4
302       CONTINUE
          CDN10(I,J)=FCDN(10.0,U10(I,J))
          OBML(I,J)=L
          RI10=10.0/L
          CE10(I,J)=FCE(CDN10(I,J),RI10)
          CH10(I,J)=FCH(CDN10(I,J),RI10)
C
C     EVAPORATION IS CALCULATED TO GIVE THE LAKE LEVEL CHANGE IN
```

```
C     M/S
C
      EVAP(I,J)=1.12*QMQS(I,J)*(UTH(I,J)-USTAR)*CE(I,J)
       HEAT(I,J)=1.12*240.*CH(I,J)*U*(TTH(I,J)-TW(I,J))
      CD10(I,J)=(USTAR/U10(I,J))**2
         IF(M.GT.20.OR.N.GT.20)GO TO 101
       GO TO 102
 110    CONTINUE
C
C      IF THERE ARE ERRORS, OR MISSING DATA EVAPORATION, TRANSFER
C      COEFFICIENTS, ETC. ARE SET TO 0.
C
       EVAP(I,J)=0.0
       CE(I,J)=0.0
        UTH(I,J)=0.0
       CE10(I,J)=0.0
       OBHL(I,J)=0.0
        U10(I,J)=0.0
       CDN(I,J)=0.0
       CD(I,J)=0.0
         CH(I,J)=0.0
       HEAT(I,J)=0.0
        RI(I,J)=0.0
            CD10(I,J)=0.0
       CDN10(I,J)=0.0
       GO TO 102
C
```

```
C       IF FOR SOME REASON THE ITTERATION DID NOT CONVERGE THEN A
C       PRINTOUT RESULTS.
C
 101     CONTINUE
        WRITE(7,500)N,M,I,J
 500    FORMAT(' NON CONVERGENT FOR N=',I3,' M=',I3,
     1 ' AT (',I3,',',I3,')')
 102      CONTINUE
          RETURN
         END
C

C       FUNCTION TO COMPUTE THE NEUTRAL DRAG COEFFICIENT USING CHARNOCK'S EQUATION
C
           FUNCTION FCDN(Z,U)
       COMMON /BLK1/USTRAN,CHARNC,CSM,NU
        IF(U.LT.0.05)U=0.05
       USTAR=.04*U
          DO 1 I=1,20
         UT=USTAR*((ALOG(9.8*Z/(USTAR*USTAR))/.4)+CHARNC)
         IF(ABS(UT-U).LT.0.05)GO TO 2
          USTAR=(U/UT)*USTAR
 1          CONTINUE
 2         CONTINUE
          IF(USTAR.GT.USTRAN)GO TO 3
           DO 4 I=1,20
          UT=USTAR*((ALOG(USTAR*Z/NU))/.4+CSM)
```

```
          IF(ABS(UT-U).LT.0.05)GO TO 3
           USTAR=(U/UT)*USTAR
4           CONTINUE
3         CONTINUE
          FCDN=(USTAR*USTAR)/(U*U)
           RETURN
          END
C
C       GIVEN THE NEUTRAL DRAG COEFFICIENT AND AN ESTIMATE FOR THE
C       RICHARDSON NUMBER COMPUTE THE ACTUAL DRAG COEFFICIENT.
C
          FUNCTION FCD(CDN,RIB)
          IF(RIB.GE.0.0)GO TO 300
         A=0.83*CDN**(-0.62)
         FCD=CDN*(1.0+(7.0/A)*ALOG(1.0-A*RIB))
         GO TO 301
300       CONTINUE
          BU=7.0
         FCD=CDN*EXP(-2.*BU*RIB)
301       CONTINUE
          RETURN
          END
C
C       COMPUTE THE BULK-TRANSFER COEFFICIENT FOR EVAPORATION
C
         FUNCTION FCE(CDN,RIB)
         IF(RIB.GE.0.0)  GO TO 400
```

```
      B=0.25*CDN**(-0.8)
       FCE=CDN*(1.0+(11.0/B)*ALOG(1.0-B*RIB))
       GO TO 401
400       CONTINUE
        BQ=20
       BU=7
       FCE=CDN*EXP(-(BU+BQ)*RIB)
401       CONTINUE
        RETURN
         END
C
C     COMPUTE THE BULK-TRANSFER COEFFICIENT FOR SENSIBLE HEAT.
C     THIS MAY BE MADE DIFFERENT FROM FCE.
C
      FUNCTION FCH(CDN,RIB)
      IF(RIB.GE.0.0) GO TO 400
      B=0.25*CDN**(-0.8)
       FCH=CDN*(1.0+(11.0/B)*ALOG(1.0-B*RIB))
       GO TO 401
400       CONTINUE
        BQ=11.0
       BU=7
       FCH=CDN*EXP(-(BU+BQ)*RIB)
401       CONTINUE
        RETURN
         END
```